Ultramafic xenoliths from alkaline basalts of Finkenberg (Siebengebirge, West Germany)

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The mantle-derived xenoliths from alkali-basalts of Finkenberg (West Germany) belong mostly to the lherzolite and wehrlite series. The members of the lherzolite suite are spinel-bearing lherzolites, harzburgites and websterites. They have undergone processes of strong deformation and partial recrystallization. The major mineral phases (olivine, orthopyroxene) are highly magnesian but the pyroxenes contain also a relatively high amount of Al_2O_3 (spinel exsolution in both ortho- and clinopyroxene). Subsolidus re-equilibration was an important stage in the formation of these rocks. This type of xenoliths is considered to represent upper-mantle material which has undergone a relatively low degree of partial melting. Xenoliths belonging to the wehrlite suite (wehrlites, clinopyroxenites) are mostly phlogopite-bearing rocks. Their magmatic textures are partly modified by later metasomatic processes. The major mineral phases are present in several generations. This feature reflects the complex multi-stage formation process of these rocks. Inclusions of the wehrlite suite are regarded as magmatic rocks, unrelated to the host basalts. The important role of such components as: K, H₂O but also Fe, Ti, Cl, and P₂O₅ in the processes of metasomatism is stressed.

Alkaline clinopyroxenites with Na-Fe-bearing green clinopyroxene, apatite, titanite, and carbonate are magmatic rocks which can not be considered members of the suites mentioned above. The important role of metasomatism processes (involving such components as Na, CO_2 and also Fe, Ti, Cl, and P_2O_5) predating the generation of the parental magma is evident.

The variety of xenolith types shows that the upper mantle beneath this volcanic region must be heterogeneous.

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Introduction

Ultramafic nodules occur in alkaline basalts throughout the world. They are widely considered as mantle-derived xenoliths and thus as an important source of information about the nature and the composition of the upper mantle as well as about the origin of alkali-rich basic magmas. Despite the great variety in composition of the nodules found in basalts, the most common rock types of the xenolith population have been established and a classification of ultramafic inclusions has been proposed.

White (1966) recognized two groups of ultramafic xenoliths: the dunite-wehrlitegabbro and the lherzolite suites. Later Kuno (1969) designated them as the lherzolite and wehrlite groups or series. These groups were defined by Wilshire and Shervais (1975) as Cr-diopside and Al-Ti-augite series (Group I and II of Frey & Prinz, 1978). The lherzolite or Cr-diopside series is widely considered to represent fragments of mantle material modified by partial melting. The members of the wehrlite or Al-Ti-augite suite were usually regarded as high-pressure cumulates precipitated from an alkali-rich basaltic magma. They may be cognate with the host basalts or accidental (White, 1966; Kuno, 1969; Irving, 1974; Wilkinson, 1975; Wilshire & Shervais, 1975; Ellis, 1976; Frey & Prinz, 1978). This classification is generally accepted but in the last few years some particular types of nodules have been reported from kimberlites and alkali-basalts which are distinct from those belonging to the series mentioned above. These are for example: the MARID suite xenoliths (Dawson & Smith, 1977), the amphibole-apatite nodules (Wass, 1979, 1980) and the metapyroxenite xenoliths (Griffin et al., 1984).

The presence of hydrous minerals (amphibole, phlogopite) in mantle-derived xenoliths, their development in veins and the presence of replacement phenomena show that processes of metasomatism are largely developed in the upper mantle and thus must be taken into consideration in the classification of mantle-derived xenoliths. The abundance of alkalis, LREE and other incompatible elements in alkaline basalts compared with those in ultramafic rocks (peridotites, lherzolites), thougt to be the source of alkalibasaltic melts, is inexplicable without the suggestion that the mantle must have been enriched in these elements before the processes of partial melting started (Oxburg, 1964; Aoki & Kushiro, 1968; Varne, 1970; Wilshire & Trask, 1971; Boyd, 1973; Best, 1974; Lloyd & Baily, 1975; Boettcher et al., 1979; Menzies & Wass, 1983). Isotopic and geochemical studies have shown that such elements as Fe, Ti, K, Na, and the LREE group as well as CO_2 and H_2O have been introduced in the mantle to form amphibole- and phlogopite-rich metasomatized ultramafic rocks. The petrographical and chemical evidence of mantle metasomatism can be considered as conclusive.

The ultramafic nodules in alkali-basalts of Finkenberg show a variety of rock-types from dunite to alkali-clinopyroxenite. These rocks differ strongly from each other not only in their composition and texture but also in their mode of formation and the role of different kinds of processes (crystallization, subsolidus equilibration, recrystallization, metasomatism) in their history. They belong to different suites and reflect the heterogeneity of the underlying mantle. Table 1 illustrates the main groups of xenoliths in a very general way.

It should be noted that all minerals have been determined by optical methods.

The specimens described in this study belong to an extensive collection that was sampled during the first decades of this century by the late Dr H. M. E. Schürmann in Finkenberg (Siebengebirge, West Germany), at a locality which is no longer accessible. This collection is currently accomodated with the Rijksmuseum van Geologie en Mineralogie (National Museum of Geology and Mineralogy) in Leiden, The Netherlands (sample numbers preceded by RGM).

Acknowledgements

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Host basalts

Basalts are aphanitic rocks with phenocrysts of olivine and clinopyroxene. Usually olivine predominates but in some samples the ratio of olivine/clinopyroxene phenocrysts is about 1:1. Phenocrysts form about 10-20% of the rock. In some samples phenocrysts have euhedral forms and are like fragments of crystals (tuffs?). A fluidal structure can occasionally be observed: bands of basalts with a different grain-size have a subparallel orientation. In a few samples (RGM 319 008, 319 080, 319 088), amygdales filled by zeolites and carbonate are present. Phenocrysts of olivine (Fo₈₂₋₈₄) are larger and more abundant than those of clinopyroxene, but olivine is absent from the groundmass. Olivine grains (\emptyset 0.3-1.0 mm) often form glomeroporphyric clots; single euhedral crystals of olivine are also common. They are often partly replaced by serpentine and carbonate. Xenocrysts of olivine are also present; they can be distinguished from the phenocrysts by their clastic habit and larger dimensions. Sometimes this olivine is more magnesian (Fo₈₆₋₈₈). The xenocrysts are strongly carbonatized.

Phenocrysts of clinopyroxene (\emptyset 0.2-0.8 mm) form single crystals as well as glomeroporphyric aggregates. Clinopyroxene is a titanium-rich augite ($2V\gamma = 56-62^\circ$; $c \not \Delta \gamma = 46-50^\circ$; r > v; twins // (100)). Zonation and dispersion are very typical; sometimes hour-glass structures can be observed. The augite has a brownish or pink-brownish colour

Suite	rock types	mineral assemblage I	pyroxene	texture	deformation crystall	ization subsolid equilibra	us metasomatisn ttion
Lherzolite	spinel-lherzolite spinel-harzburgite spinel websterite	$\begin{array}{l} Ol + Opx + Cpx + Sp \\ Ol + Opx + Cpx + Sp \\ Op + Cpx + Sp \end{array}$	Al-rich enstatite Al-rich diopside	porphyroclastic	strong	+	I
Wehrlite	phlogopite-bearing wehrlite phlogonite-bearing	OI + Cpx + PhI	Ti-rich aug	poikilitic	weak	 +	+
	clinopyroxenite	$Cpx \pm Ol + Phl$ Cpx + iron-ore	Ti-rich aug Fe-Ti-rich	coarse granular	weak weak	++	+
	apatite-bearing clinopyroxenite	Cpx + Ap + Sph + Carb	aug (salite) Fe-Na-rich aug		weak	+	+

Table 1. Xenolith types in Finkenberg basalt.

and is mantled by a more intensively coloured pinkish rim. Often the cores of phenocrysts are colourless or pale-green. In such grains the boundary of the core with the border zone can be seen only under crossed polarizers. The core of the clinopyroxene grains can have a clearly green or yellow-green colour (in some samples with pleochroism) and occassionally a spongy habit. The optical properties of such cores $(2V\gamma = 60-62^\circ; c \measuredangle \gamma = 38-40^\circ)$ are distinct from those of augite. Boundaries between the core and the pinkish rim are very sharp; the form of the core is irregular, resorbed and only the pinkish rim gives to the whole grain a euhedral form.

Comparison of this type of clinopyroxene crystals with titanaugite rims on spongy diopside grains of nodules shows that they are very similar. Also, the green clinopyroxene crystals seem to be identical to the green clinopyroxene of the clinopyroxenite nodules. Grains of the green pyroxene in the basalts are usually larger than the phenocrysts of titanium-rich augite. All these facts lead to the conclusion that green clinopyroxenes (in separate grains or yellow-green cores) have no relation to the host basalts and are fragments of xenoliths.

Xenocrysts of olivine (Fo₈₆₋₈₈) and diopside derived from xenoliths are often present. The groundmass of the basalts consists of titanaugite, microlites of plagioclase (An₅₆₋₆₄) and magnetite. Occasionally rare flakes of brown mica are present. The presence of nepheline or alkali-feldspar is difficult to establish. The composition of the groundmass varies hardly. Clinopyroxene is always the dominant mineral. The average composition of the groundmass is: 50% (40-60%) clinopyroxene, 35-40% (20-40%) plagioclase and 7-15% magnetite. Secondary minerals are: serpentine, carbonate, zeolites, and chalcedony replace the primary mineral phases or form numerous veinlets cutting basalts. They are often located along contacts with nodules.

Chemical analyses of the Finkenberg basalts (Frechen, 1948) show a relatively high content of alkalis (5-6%) as well as the presence of normative nepheline (6%). These rocks are thus alkali-olivine-basalts.

The basalts contain 'xenoliths' of cumulates of basaltic composition formed by the same major mineral phases and having a clear cumulus texture.

COGNATE CUMULATES (AUTOLITHS)

These rocks (RGM 319 347, 319 386, 319 403) are composed of olivine and clinopyroxene with a small amount of trapped liquid or intercumulus phase, consisting of labradorite, iron ore and pyroxene. Their modal composition varies considerably and therefore is given here separately for each sample:

RGM 319 347 – 10% Ol; 89% Cpx; 1-2% trapped liquid.

RGM 319 403 – 35% Ol; 50% Cpx: 12-15% intercumulus phase.

RGM 319 386 - 55% OI; 45% Cpx; 0.5-0.8% trapped liquid.

In sample RGM 319 347 the cumulus phase is represented by olivine. Its small (\emptyset 0.5-0.6 mm) euhedral crystals are included in a megacryst (25-50 mm) of clinopyroxene; only 2-3 other grains of pyroxene (\emptyset 2.0 mm) are present. Olivine is usually surrounded by trapped liquid or carbonate. In other samples olivine ($2V\gamma = 92-94^{\circ}$; Fo₇₈₋₈₀) forms two kinds of grains: 1) small (\emptyset 0.1-0.2 mm) rounded grains, poikilitically included in clinopyroxene or located between its grains, and 2) rounded and euhedral crystals (\emptyset 1.0-1.2 mm) of the same size as grains of clinopyroxene. Large grains of olivine contain inclusions of trapped liquid which is now represented by an aggregate of microlites of plagioclase and small grains of pink titanaugite.

Crystals of pyroxene may be enclosed by olivine and vice versa. Olivine is partly or completely replaced by carbonate and chalcedony. In some instances two types of olivine may be observed, differing from each other mostly by the degree of alteration. The one is replaced by secondary minerals and the other is practically unaltered. The first may be included in the second. It is here assumed that these different kinds of olivine represent xenocrysts and cumulus phases respectively.

The clinopyroxene is a titan-rich augite ($2V\gamma = 56-60^\circ$; r > v; $c \not = 42-44^\circ$). Its brownish or pink-brownish grains (Ø 1.5-2.0 mm) are usually euhedral or rounded. The augite exhibits a zonation and occasionally an hour-glass structure. Poikilitic inclusions of olivine and iron-ore are comon, as are inclusions of trapped liquid. Abundant cavities filled by carbonate and labradorite, occasionally along with single flakes of brown mica and isotropic material (glass or opal), also occur giving the clinopyroxene grains a spongy habit. In two samples several grains of augite contain inclusions of resorbed grains (\emptyset 0.1-0.3 mm) of a colourless clinopyroxene ($2V\gamma = 62-64^{\circ}$), probably diopside, with an altered and spongy surface. These inclusions are usually surrounded by a 'ring' of small rounded grains of olivine partly carbonatized and also included in augite. Grains of diopside can be mantled by a reaction zone of a finely dispersed material of high birefringence. A certain type of zoning can be observed: resorbed grains of diopside \rightarrow reaction zone \rightarrow small grains of olivine, all zones being inclosed by the augite. More rarely augite overgrows diopside without any reaction zone. Such relations indicate that the augite has crystallized on fragments of diopside; the crystallization may have taken place near the bottom of the magma chamber or near the walls of the channel.

Inclusions of trapped liquid in augite have different forms and sizes (\emptyset 0.05-0.8 mm). They are composed of microlites of labradorite and grains of magnetite; carbonate is often present. A pink rim mantling clinopyroxene grains in contact with the trapped liquid gives them an euhedral form. The trapped liquid differs from the veinlets of basalt cutting the nodules by the absence of titanaugite, which forms here only rims on large crystals of pyroxene.

In most cases the intercumulus phase is absent and the rocks are adcumulates (Wager et al., 1960). Only in one sample it fills the space between grains of olivine and pyroxene. The intercumulus phase is formed by microlites of labradorite (An₅₆₋₆₄), small grains of iron-ore and titanaugite. Occasionally titanaugite encloses microlites of plagioclase forming a subophitic texture. The same purplish, low-pressure augite forms rims on the cumulus titan-augite indicating that the growth of rims and the crystallization of the intercumulus liquid occurred simultaneously. Grains of magnetite (\emptyset 0.01 mm) are usually localized near the boundary of augite crystals with the intercumulus phase. Rare flakes of brown mica and thin needles of pale-green amphibole may be present, carbonate and serpentine are common. In contact with the host basalt a thin rim of purplish titanaugite around the clinopyroxene of the nodules can be present.

Typical cumulus textures indicate that the rocks described are magmatic rocks formed during a process of fractional crystallization and settling of crystals, forming mostly adcumulates (Wager et al., 1960; Jackson, 1961). The sequence of crystallization in the cumulates, as well as in the basalts, was: olivine (\pm clinopyroxene) \rightarrow clinopyroxene + plagioclase. Labradorite never forms a cumulus phase and is absent as phenocrysts in the basalts; olivine is absent from the groundmass. The similarity in composition of the mineral phases of the cumulates and basalts indicates that they have had the same parental magma. The fractional crystallization which has led to the formation of the cumulates appears to have operated at higher pressures than the formation of the basalts. The spongy habit of the augite (conversion of pyroxene in response to changing physical conditions

related to the ascent of magma to the surface) and the formation of rims of low-pressure titan-augite are evidence in support of this suggestion. The formation of cumulates is believed to have operated near the bottom of the magma-chamber (presence of xenocrysts of more magnesian olivine and of diopside included in augite).

Xenoliths of the lherzolite suite

Xenoliths of the lherzolite suite are represented in Finkenberg only by spinel-bearing rocks (Table 2). Spinel-bearing ultramafic xenoliths represent the most common type of nodules recorded from different areas of alkali-basaltic volcanism all over the world. The most widespread type of rocks – spinel-lherzolite – was found in basalts differing in composition from alkali-olivine-basalts to the extremely alkali-rich varieties such as nephelinites, or leucitites.

Petrological, chemical and geochemical data have shown that there are no genetic relationships between xenoliths and their host lavas. Therefore it is generally accepted that spinel-lherzolites are accidental xenoliths and must represent upper mantle material.

The ultramafic nodules of the lherzolite suite of this study are mostly spinel-bearing lherzolites but harzburgites and websterites, as well as transitional types, are also present. The rocks consist of varying proportions of minerals of the four-phase spinel-lherzolite assemblage: Ol + Opx + Cpx + Sp. Only orthopyroxene and spinel are present in all types of rocks. An important feature of this group of inclusions is a considerable degree of deformation with the development of porphyroclastic textures. Often relics of poikilitic textures can be observed.

LHERZOLITES

The modal composition of the lherzolites (RGM 319 307, 319 348, 319 378, 319 392, 319 404, 300 000) is: 45-50% (35-57) olivine, 20-30% (15-50) orthopyroxene, 20-30% (10-37) clinopyroxene, and 1-5% spinel. The commonly observed texture is porphyroclastic. In two samples relics of a poikilitic texture can be noticed (Pl. 1, figs. 1, 2). In sample RGM 319 404 a heterogeneous distribution of minerals has been observed: one part of the nodule consists of 60-100% of olivine with subordinate pyroxene, the other one mostly of pyroxene. The boundary between these two parts is sharp but not obviously intrusive.

Olivine — It is a forsterite (Fo₈₈₋₉₂; $2V\gamma = 86-90^{\circ}$). Three types of olivine grains are present: 1) small grains (\emptyset 0.2-0.4 mm), partly or completely included in orthopyroxene (relics of a poikilitic texture); 2) porphyroclasts of a different size (\emptyset 1.0-6.0 mm) and 3) grains (\emptyset 0.5-1.5 mm), partly or completely recrystallized. Olivine included in orthopyroxene is often replaced by serpentine, carbonate and talc and shows no signs of deformation.

The amount of porphyroclasts in different nodules is variable; sometimes only 3-4 are present, in others they are so numerous that the matrix forms only 10-15% of the rock. The porphyroclasts are often enlongated and show a subparallel orientation. They have irregular curvilinear outlines and show strong deformation. Undulose extinction, cracks, subgrains, deformation lamellae (parallel to (100)), and kink-bands are common. Occasionally porphyroclasts contain zones and bands of micro-inclusions which are mostly secondary. The olivine porphyroclasts are cut by thin veinlets of serpentine, talc and carbonate.

xenoliths.
ultramafic
Phlogopite-bearing
Table 4.]

			a post														
Samp le	Bock	БО	1 sod	tlon	Texture	Defor-	ч	aracteristics of	minerals			-	0 1				
R.G.H.		10	Срх	۱۹۹		mation	10	Срх	ГЧd	Liquid/solid	Metasomatism	Reactions	Deformat recrysta	ion Re-equilibrat	tion met	Contact tamorphism	Other processes
319 396	Vehrlite	45	0 7	÷	polkilitic (banding)	weak	Fo ₈₀₋₈₂ 1, 11	Salite (?) ,	I. 11b, 111, IV	01 I + Phi I 01 II + Cpx I	Phi lib+CpxI				Cpx Ph1	x 1+Cpx 1+Ph1 V	Phl I +Phl I + Ap + Mgt
319 038	Vehrlite	55	30	15	poikilitic	Yea X	Fo 80-82 1, 11	Salite (aug 7) 1, 11	1, 11a, 111	0 + Mgt + Ph 0 + Cpx	Ph1 - La			Cpx 1+Cpx 11	+ + L		
319 139	Vehr1îte	50	35	15	poikilitic (banding)	weak	<u>۔</u> 	Salite (aug ?) 1, 11	1, 11b, 111, 1V	01 1 + Ph1 1 01 1 + Cpx 1	Phi i i bCpx i (weak)			Cpx I+Cpx I II+	L+Sp7 Ph1		Phi II-Phi III + Ap
319 390	Vehrlite	75	22	~	poikilitic	weak	^{Fo} 82-84	Salite (aug ?) 1, 11a, 111	-, ia, ii	01 I + Phi I 01 I + Cpx I Cpx I + Phi IIa	Phl Lia			Cpx 1-Cpx 11 Cpx 1-Cpx 11	+ L Cpx + Jd? Cpx	I+Cpx +Cpx	Phi II+Phi III + Ap
319 394	Vehrlite	55	35	01	porphyroclastic relict poikilitic	strong	Fo ₈₃₋₈₅ 1, 111, 1V	Salite (?) 1, 11, 1V	11b, 111	01 01 12 + Cpx	Phi I I b Cpx I		01 CP	×			
319 025	Vehr ī te	50	45	<u>~</u>	porphyroclastic relict polkilitic	strong	Fo ⁸⁰ 1, 111, 1V	Salite (?) I, IV	1, 11b, 111, IV	01 I + Phi I 01 I 1 + Cpx I	Ph b-+ Cpx		01 Cp	x? Cpx I+Cpx II	+ L Cpx	111Phi IV c 1Cpx 111	Phi I (++Phi I (+4 + Ap
319 397	Wehrlite	45	07	15	porphyroclastic relict poïkilitic	strong	Fo ₈₀ 1, 111, 1V	Salite (?) 1, IV	I, 11b, 111	0 + Phl 0 7 + Cpx	Ph1 11b Cpx 1. (weak)		0J	×1			
101 616	01+clīnopyroxenite	2	60	30	poikilitiç	weak	Fo80-81 1, 11, V	Augite (7) , 111	1, 11a, 111, 1V	01 + Mgt + Phl Cpx + Phl a	Phi Ia	Ph1 + 0; 01 V + L			Ph1	111	
319 088	Clinopyroxenite	۳.	8	17	coarse granular	weak	Fo ₈₂₋₈₄ v	Augite (?) , ,	111, 111	Cpx 1	Phi 11b-+ Cpx (weak)	Ph1 + Di 01 v + L		Cpx I+Cpx II	+ L Cpx	k I-=Cpx Cpx	
319 080	Clinopyroxenite		70	30	coarse granular	weak		Augite (?) 1, 111	l, 1a, 1	Рні I Срх I + Рні IIa	Ph1 1a				Cpx	4 1-Cpx 111	
319 407	Dunite	85		15	coarse granular		^{Fo} 86-887 I (relics)		116, 111	10	Ph1 115						
								GEWER	ATIONS 0	FMINER	ALS						
		poiki separ porph neobl react	liti. ate (yroci asts ional	cally grain: lasts lgrai	included in Cpx I s ns			Cpx I - 01koc 9ra1r. 11 - broke 11? - 10w-p IV - neob1	rysts in wehrlit. Is with microincl In down Cpx (part Pressure Ti-rich asts	es and Ol-clin. usions in clinx jal melting) augite-rims	opyroxeni tes; opyroxeni tes		 I4	<pre>1 - poik litical 1 a - "primary-met 1 a - "secondary-m 1 b - "secondary-m 1 - "late metaso 1 V - contact-meta</pre>	ly include asomatic' tetasomatic' matic'' (se morphic (T	ed In Cpx c'' scondary) fi-rich)	

Two types of olivine grains are present in the matrix: a) strongly deformed grains, and b) neoblasts, forming a granoblastic aggregate with triple point junctions as grain boundaries at about 120°. Both fill the interstices between porphyroclasts.

Orthopyroxene — It is an enstatite or enstatite-bronzite $(En_{88.92})$ ($2V\gamma = 84-90^{\circ}$) with a relatively high content of Al_2O_3 (exsolution lamellae of spinel). It occurs as porphyroclasts (\emptyset 3.0-7.0 mm) and small grains (\emptyset 0.5 mm) forming the matrix. The form of the porphyroclasts is mostly irregular and occasionally ellipsoidal with curvilinear boundaries. They are strongly deformed and undulose extinction, subgrains and kink-bands are wide-spread.

The orthopyroxenes contain numerous inclusions of different kinds: a) small grains of olivine mentioned above and rarely small grains of brown spinel; b) translucent and opaque micro-inclusions forming linear zones and bands. These inclusions are primary as well as secondary, melting cavities are probably also present. The enstatites contain exsolution lamellae of clinopyroxene oriented parallel to (100) of the host grain. In some samples (RGM 319 348) the clinopyroxene is recrystallized and forms broad lamellae; it can also replace the host orthopyroxene along the cleavage planes or form rims around enstatite grains. The orthopyroxene contains numerous exsolution lamellae of spinel which are present in all lherzolites with a porphyroclastic texture. The form of the spinel lamellae is variable; in some sections they appear as long continuous needles, in others as short bars or rods, the most common are very thin blades, plates, or short needles. The exsolution lamellae are oriented in one or two directions, often forming a sort of grid. In differently oriented sections of orthopyroxene the angle between the spinel lamellae directions may change from 45° to 90° (Pl. 4, fig. 1). The lamellae can be oriented parallel to (100) and (010), or parallel to [001] within (010), and sometimes parallel to [100] in (010); in one sample they seem to be parallel to (001). Some exsolution lamellae are recrystallized in blebs or small grains.

The orthopyroxene porphyroclasts are cut by veinlets of secondary minerals such as talc and carbonate. Enstatite is often mantled by a turbid zone of finely dispersed material of high birefringence. These zones can be as wide as the grain they surround (RGM 319 348, 319 378). The same finely dispersed material can be seen filling cracks in orthopyroxenes. At the contact with the host basalt the turbid zones around the enstatite are always present. Small grains of orthopyroxene in the matrix are not very numerous, some of them are recrystallized and form a mosaic of grains with triple point junctions at 120° (RGM 319 005, 319 404), others are fragments of deformed large grains.

Clinopyroxene — This is a colourless or pale-green diopside $(2V\gamma = 56-62^\circ; c \pm \gamma = 39-41^\circ)$ containing Al₂O₃ (exsolution lamellae of spinel), Na₂O (plagioclase in zones of partial melting of the clinopyroxene) and probably Cr₂O₃. In rare samples (RGM 300 000) twins parallel to (100) have been observed. Occasionally the clinopyroxene forms large grains (Ø 3.6-5.0 mm) containing numerous inclusions of rounded olivine grains. One nodule (RGM 319 404) contains clinopyroxene porphyroclasts, but usually the diopside forms small anhedral grains (Ø 0.5-0.8 mm) interstitial to both olivine and orthopyroxene and tends to occur as aggregates of four to six or more grains (RGM 319 348, 319 392, 300 000). Occasionally such grains are deformed (undulose extinction and subgrains). The diopside contains besides olivine grains small grains of iron-ore and brown spinel and numerous micro-inclusions. Very typical is the presence of exsolution lamellae of spinel, oriented in different directions. Thin blades or needles of pale-green spinel are parallel to (100) or to [001] in (010). In differently oriented sections of the diopside the angle between the needles of spinel may vary from 45 to 85°. The size of the clinopyroxene aggregates varies from \emptyset 0.4 to 4.0 \times 4.0 mm. Occasionally a sort of zonation can be observed: a concentration of small grains of ore or its intergrowth with diopside occurs in the centre of such an aggregate (Pl. 1, fig. 1); in this part also numerous small grains of brown spinel are included in the diopside. In the peripheral parts, clinopyroxene contains inclusions of olivine. The clinopyroxene grains have often a spongy border zone or even the whole grain may have a spongy habit (Pl.2, fig. 2). Often pyroxene grains are divided in sections whilst preserving their uniform optical orientation and birefringence (Pl. 3, fig. 1). The interstices between sections are filled by microlites of plagioclase, small grains of brown spinel, opaques, serpentine, carbonate, and chalcedony; flakes of phlogopite are rarely present (RGM 319 348). In spongy zones plagioclase and spinel as well as carbonate are often present. The ubiquitous association of clinopyroxene with plagioclase and spinel in spongy zones or spongy grains (generally regarded as a result of partial melting), or with the 'desintegrated' grains of diopside, seems to indicate the breakdown of diopside into plagioclase and spinel according to the equation: $Cpx_1 \rightarrow Pl + Sp + Cpx_2$. Often a rim of pink titanium-rich augite is formed around diopside with spongy border zones or around the small sections of 'desintegrated' grains.

The form of the clinopyroxene aggregates may be irregular and/or isometric, but often they have the form of veinlets penetrating between the olivine grains. Such a form leads to the suggestion that these aggregates became mobile during the process of partial melting (Pl. 2, fig. 1). Some grains of clinopyroxene are recrystallized and these neoblasts (\emptyset 0.1-0.2 mm) fill cracks in large grains of orthopyroxene.

Spinel — It is usually a deep-brown or red-brown picotite, sometimes with a greenish or pinkish tint. A zonation can be locally observed: a brown isotropic core is mantled by a greenish zone of low birefringence. Occasionally the border zone of spinel grains is converted into an opaque iron-ore.

The spinel shows three modes of occurrence: 1) exsolution lamellae in pyroxenes, 2) minute grains included in clinopyroxene, and 3) separate irregular, rounded, or vermicular grains (\emptyset 0.2-0.4 mm) interstitial to silicates. Often the spinel forms long stringers penetrating between silicate grains or along the cleavage planes or cracks in pyroxenes (Pl. 3, fig. 2). This mobility of spinel is an important feature.

Grains of spinel may be surrounded by aggregates of serpentine or carbonate. Spinel exposed on the nodule margins is converted into an opaque iron-ore.

HARZBURGITES

This group (RGM 319 163, 319 908, 319 005) includes rocks consisting of olivine, orthopyroxene and less than 10% clinopyroxene (Streckeisen, 1976). Such rocks are often regarded as lherzolites. It is rather difficult to distinguish them from lherzolites containing only 10-12% clinopyroxene (RGM 319 378, 319 404). There are no differences between lherzolites and harzburgites in texture or characteristics of minerals.

The average modal composition of harzburgites is: 50% olivine, 35-40% orthopyroxene, 7-8% clinopyroxene, and 2-5% spinel. One sample contains 72% olivine, but it may represent part of a harzburgite or lherzolite nodule with banding.

All rocks are strongly deformed and have a porphyroclastic texture with a small amount of porphyroclasts and with relics of a poikilitic texture. The degree of recrystallization is rather low. Olivine — It is represented by forsterite (Fo₈₈₋₉₀; $2V\gamma = 87-90^{\circ}$) and occurs in three forms: 1) as small grains (\emptyset 0.2-0.8 mm) poikilitically included in orthopyroxene; 2) as porphyroclasts (\emptyset 4.0-8.0 mm) and 3) as grains filling spaces between porphyroclasts. The first two forms are quite identical to those of the lherzolites (described above). The grains filling spaces between porphyroclasts are strongly deformed. Neoblasts are rare.

Orthopyroxene — This is an enstatite $(En_{87-90}; 2V\gamma = 83-90^{\circ})$ with a high content of Al₂O₃ (exsolution lamellae of spinel). It forms porphyroclasts (Ø 3.0-8.0 mm) and small grains in the matrix. The outline of the porphyroclasts is related to the presence or absence of olivine inclusions in their border zones. The presence of these inclusions gives them a curvilinear contour. Deformation (cracks, microfractures) led to a different type with a complex boundary line. These two different forms illustrate the coexistence of two types of textures – poikilitic and porphyroclastic. Some porphyroclasts are elongated, others isometric. The orthopyroxenes are always strongly deformed with undulose extinction, bending of grains, subgrains, etc. They contain inclusions of olivine, occasionally spinel and micro-inclusions forming linear zones. Exsolution lamellae of clinopyroxene in enstatite are fine and straight; occasionally they are curved along with the host grain. Exsolution lamellae of clinopyroxene and spinel are present in one and the same grain of orthopyroxene, they may have a subparallel orientation (// to (110)).

Near veinlets of secondary minerals and in the border parts of enstatite grains the exsolution blades of spinel are recrystallized and have become blebs and lobate grains. Small grains of orthopyroxene are recrystallized and fill spaces between porphyroclasts or fine cracks in large grains of enstatite.

Clinopyroxene — The clinopyroxene of the harzburgites is identical to the diopside of the lherzolites and has the same particuliarities – spongy rims, exsolution lamellae of orthopyroxene and spinel, etc.

Spinel — Deep-brown spinel grains (\emptyset 0.3 mm) often have a vermicular habit and fill spaces between silicate grains.

WEBSTERITES

The average modal composition of these rocks (RGM 319 374, 319 391) is: 70-75% clinopyroxene, 20-25% orthopyroxene, and 3-5% spinel. The texture is porphyroclastic with numerous porphyroclasts of orthopyroxene and clinopyroxene (Pl. 5, fig. 1). The matrix is formed mostly by clinopyroxene but also by orthopyroxene and spinel. The rocks are strongly deformed and partly recrystallized.

Orthopyroxene — This is an enstatite (En_{86-88}) close to bronzite ($2V\gamma = 84-90^{\circ}$) with a relatively high content of Al_2O_3 (exsolution lamellae of spinel). Three forms of grains are present: 1) exsolution lamellae in clinopyroxene, 2) porphyroclasts and 3) neoblasts. Exsolution lamellae are straight, fine and continuous. In one rock (RGM 319 391) they form large, 1.0 mm wide bands, rather like twins or intergrowths. Such bands are localized not only within grains but also between them. All bands of orthopyroxene are optically uniformly oriented. This is possibly a result of recrystallization. Porphyroclasts of orthopyroxene ($\emptyset 2.0-3.0$ mm) have an irregular form and curvilinear boundaries. Occasion-

ally they are elongated but a linear orientation is practically absent. The orthopyroxenes are strongly deformed and cracks, subgrains and kink-bands are common.

Exsolution lamellae of spinel in orthopyroxene are oriented in one (// to (100)) or two (// (100) and to [100] in (010)) directions. They may be parallel to exsolution lamellae of clinopyroxene. Spinel forms occasionally veinlets cutting through orthopyroxene grains. The enstatites contain microinclusions of different kinds – primary as well as secondary. As usual they form zones and bands. Occasionally the orthopyroxenes have spongy border zones with abundant cavities of melting. Like orthopyroxenes in other rocks, its grains are mantled by a turbid zone of finely dispersed material of high birefringence. Small grains of plagioclase, carbonate and olivine can be distinguished in these zones. The turbid zone is especially wide at the contact with the host basalt.

The amount of neoblasts (\emptyset 0.2-0.4 mm) of orthopyroxene in the matrix is rather small; they fill cracks in porphyroclasts and surround the large grains.

Clinopyroxene — It is a colourless or very pale-green diopside $(2V\gamma = 62-64^\circ, c \pm \gamma = 38-42^\circ)$ rich in Al₂O₃ and Na₂O. It forms porphyroclasts and neoblasts in the matrix. The porphyroclasts, ranging in size from 1.2×2.8 mm up to 1.6×4.0 mm, have an irregular form and are strongly deformed. Zones of strain, undulose extinction and subgrains are common. Exsolution lamellae of orthopyroxene are curved along with the host grains. Exsolution lamellae of spinel are always present and oriented in one or two directions. In section \pm [001] exsolution lamellae of spinel show an angle of c. 42° with the cleavage-planes (Pl. 4, fig. 2). An interesting type of exsolution lamellae of orthopyroxene. In this case needles of spinel are localized in the exsolution lamellae of orthopyroxene and are oriented parallel to each other (within the lamellae of orthopyroxene) but make an angle of 45-50° with the direction of these lamellae. They are very numerous inside the lamellae of orthopyroxene but rare outside them. The numerous inclusions and cavities in the diopside give to the whole grains a spongy character. Spongy zones are also present and partial melting and breakdown of clinopyroxene can be observed.

The diopside exposed at nodule margins is mantled by Ti-rich pinkish augite. Occasionally augite forms an uninterrupted band along the whole nodule.

Neoblasts of clinopyroxene form a granoblastic aggregate surrounding the porphyroclasts and fill small cracks in large grains of diopside.

Spinel — Greenish brown spinel forms lobate grains (\emptyset 0.8-1.0 mm) of an irregular or vermicular form, filling the space between grains of pyroxenes (Pl. 5, fig. 2). Minute grains of spinel are always spatially associated with aggregates of 'desintegrated' diopside.

Plagioclase — Few grains of plagioclase (andesine-labradorite) are present in websterites. They form large xenomorphic grains interstitial to pyroxenes.

The lherzolite (harzburgite-websterite) nodules of Finkenberg are very similar to those described from other areas of the world. They have a relatively constant mineral composition and belong to the most common lherzolite (or Cr-diopside) suite of ultramafic xenoliths. A diagnostic feature is the high magnesium content of their major mineral phases (Fo₈₈₋₉₂, En₈₇₋₉₂) as well as the relatively high Al_2O_3 -content of their pyroxenes (exsolution lamellae of spinel).

The clearly metamorphic, porphyroclastic texture of the rocks indicates that they have undergone strong deformation processes. Only in three samples the texture can be considered as coarse granular; in all other rocks the presence of porphyroclasts and of a partly recrystallized matrix leaves no doubt of the importance of the processes of deformation and recrystallization for the formation of these rocks. Nevertheless, relics of poikilitic textures show that the lherzolites have had a pre-deformation history related to solid/liquid processes: processes of crystallization. For some rocks the following order of crystallization may be suggested:

$(Sp + iron-ore) \rightarrow Ol (included in pyroxenes) \rightarrow Opx \rightarrow (Opx + Cpx).$

It may be assumed that the crystallization took place under high-pressure conditions, because of the high Al_2O_3 -content in the pyroxenes. The crystallization was followed by a subsolidus stage. The widely developed exsolution processes (exsolution of clinopyroxene from orthopyroxene and vice versa, exsolution of spinel from pyroxenes) are typically related to the subsolidus equilibration. Thus, the evidence for an early stage of high-pressure and high-temperature in the formation of the lherzolites is preserved in the form of poikilitic texture and exsolution processes in pyroxenes. The subsequent stage of deformation and related recrystallization (porphyroclastic texture, recrystallization of the matrix) seems to have occurred also before the incorporation of the lherzolites into the host basalts. The deformation of exsolution lamellae is one of the evidences for such a succession of processes.

It is difficult to establish the time of these processes of recrystallization and reaction, such as the recrystallization of exsolution lamellae of pyroxene and spinel, the formation of clinopyroxene rims on enstatite, and the recrystallization of spinel. All these processes can have occurred during a long period of time.

After the incorporation of nodules in the ascending basaltic magma, the change of P-T conditions had led to another sequence of important processes. First the pyroxenes had melted partially. Spongy borders and spongy grains of diopside with the formation of glass, and the breakdown of clinopyroxene in spinel, plagioclase and a new pyroxene resulted from this change of conditions. Incongruent melting of orthopyroxene is less evident but zones of fine-grained material containing olivine could be the result of this process. During this stage, some parts of the rocks became very mobile and subject to plastic flow (vein-like aggregates of diopside, vermicular grains and veinlets of spinel). The presence of newly formed plagioclase in association with clinopyroxene in places of partial melting indicates that these processes occurred under relatively low pressures.

The last important stage in the formation of the lherzolites was their reaction with the host basalts. These processes have not strongly changed the texture nor the character of the mineral phases of the rocks, but they show that the basalts and the xenoliths were not in equilibrium, and thus were formed under different physical conditions. The reactions at the contact of the rocks in question are: the formation of iron-ore rims on spinel, rims of fine-grained material on orthopyroxene grains, and the low-pressure titanaugite mantling the clinopyroxene.

Thus, the petrographic study permits to establish a complex multi-stage process of formation of ultramafic inclusions before and after their incorporation in basalts. In a very general way the sequence of these processes can be presented as follows:

crystallization \rightarrow subsolidus re-equilibration \rightarrow deformation and related recrystallization \rightarrow incorporation in basalt \rightarrow partial melting \rightarrow contact metamorphism.

DISCUSSION AND CONCLUSIONS

Spinel-bearing ultramafic xenoliths of the Cr-diopside or lherzolite suite are recorded from basalts of all areas with continental alkali-basaltic volcanism. They are the most abundant and widespread type of mantle-derived nodules found in basalts. The mineralogical and

chemical composition as well as the geochemical affinities of lherzolites from different localities are very similar. The most common are olivine-rich lherzolites, but harzburgites, websterites and clinopyroxenites also occur. The relations between these different types can be observed in complex xenoliths, and in most cases pyroxenites form veins (clinopyroxenites) or bands (orthopyroxenites) in olivine-rich rocks.

The modal mineral assemblage of the dominant rock type is: 65-75% OI + 20-30% Opx + 8-12% Cpx + 1-2% Sp. The composition of the major mineral phases is also uniform for most localities. The olivine is usually a high-magnesian forsterite; the orthopyroxene is an enstatite or enstatite-bronzite with a high Mg/ Σ Fe ratio and a relatively high Al₂O₃-content (1.7-3.9%; cf. Cooper & Green, 1969; Donaldson, 1978), established by direct analyses but also based on the presence of exsolution lamellae of spinel. The clinopyroxene is a Cr-diopside or endiopside. It contains also Al₂O₃ and occasionally Na₂O (White, 1966; Wilkinson, 1973; Frey & Green, 1974; Francis, 1976). Exsolution lamellae of spinel in pyroxenes are widespread. These rocks are often thought to be 'a subsolidus assemblage'.

The olivines, orthopyroxenes and clinopyroxenes are richer in MgO and Cr_2O_3 , when compared with those of the wehrlite series. The high Al_2O_3 -content is considered evidence for the high-pressure crystallization of the lherzolites. Such a mineral assemblage is stable (as documented by experimental work-systems CMAS and MAS) over a wide P-T range. The absence of garnet and plagioclase indicates that the formation of these rocks was limited by a pressure interval from 8 to 18 kb.

The composition of spinel-bearing xenoliths of this study is very similar to those described elsewhere. The extremely uniform composition of spinel-lherzolite nodules found in basalts with very different chemical characteristics (from olivine-alkali-basalts to phonolites or leucitites) shows that the ultramafic rocks cannot be considered cumulates precipitated by such different parental magmas. For the same reason they cannot just reflect some specific regional processes but rather must be related to processes characteristic for alkali-basaltic volcanism on a world-wide scale. Such a process must be the partial melting of the upper mantle and the generation of basaltic magmas. This suggestion is confirmed by the ubiquitous association of xenoliths with basaltic lavas. Thus, ultramafic xenoliths can provide information about the composition of the upper mantle, the processes of its partial melting and probably also about the P-T conditions of these processes (Green & Ringwood, 1967; Forbes & Kuno, 1967; Kuno, 1969; Carswell et al., 1979; Carswell, 1980; Kuno & Aoki, 1970; Carter, 1970; Frey & Green, 1974; Frey et al., 1978; Wilkinson, 1975; Maaløe & Green, 1977; Ghent et al., 1980; Dawson, 1980). It is also of great interest whether the ultramafic rocks are fragments of primordial mantle material or of a mantle depleted during processes of partial melting. They could also be residues left after the extraction of basaltic melts. Some criteria based on petrological, chemical, geochemical, and experimental studies to distinguish these different types of mantle material have been proposed (Kleeman et al., 1969; 1970; Kushiro, 1969; Trask, 1969; Paul, 1971; Ridley & Dawson, 1975; Meysen & Kushiro, 1977; Stosch & Seck, 1980; Kurat et al., 1980; Kyser et al., 1982; Takashi & Kushiro, 1983; Stolz, 1984). Detailed petrographic studies have shown the existence of close relationships between the equilibrium conditions of ultramafic rocks, their composition and their textures.

Many authors have studied the texture of lherzolitic xenoliths which is usually regarded as a texture of 'metamorphic tectonites' (den Tex, 1969; Pike & Schwarzman, 1977). Bouillier and Nicolas (1973, 1975) have distinguished different types of textures of xenoliths in kimberlites: coarse granular and tabular textures representing a 'rigid' mantle or mantle material deformed at a 'very slow strain rate', and porphyroclastic and mosaic

textures formed under conditions of plastic flow and recrystallization. Boyd and Nixon (1975) have introduced the terms 'granular' and 'sheared' lherzolites. According to these authors the granular textures originated in the depth-range of 100-150 km and granular lherzolites are depleted in Na, Ti, Fe, Ca, and Al relative to the sheared lherzolites which are strongly deformed at depths of 150-200 km, and enriched in these elements. This difference in composition is considered by Boyd and Nixon as a result of varying degrees of depletion. Ehrenberg (1982) prefers to relate the enrichment in Fe and Ti of the sheared rocks to a process of metasomatism. Similar classifications were proposed by Harte (1977) for garnet-bearing lherzolites and by Basu (1977) for spinel-bearing ultramafic nodules from basalts. These classifications differ from each other mostly in details. Granular – coarse-granular – coarse texture of different workers are synonyms, as well as sheared and porphyroclastic textures. The first group is characterized by weak deformation (strain, undulose extinction, subgrains) of olivine, exsolution lamellae of clinopyroxene in or-

thopyroxene, a vermicular form of spinel grains, and the absence of recrystallization. The porphyroclastic texture is characterized by strong deformation and by two generations of olivine, clinopyroxene and orthopyroxene (porphyroclasts and neoblasts) and four generations of spinel (Basu, 1977). Exsolution lamellae of various kinds are widespread. This texture shows evidence of strong plastic deformation and accomodation by recrystallization. According to Basu (1977) the degree of deformation decreases with depth and the most deformed rocks are depleted in Na, Al, Fe and enriched in Cr en Ni; thus a trend different from that described by Boyd and Nixon for xenoliths from kimberlites (sheared rocks are enriched in Fe, Na and Al).

The completely recrystallized textures are: mosaic, equigranular and mosaic-tabular. It is often difficult to estimate the relative role of metamorphic and igneous processes or processes of partial melting and metasomatism in the formation of the textures of lherzolites, but some considerations have been proposed.

For the ultramafic nodules from basalts it is generally accepted that more deformed rocks are more depleted in the low-melting fraction and represent a more refractory mantle material than coarse or coarse-granular ultramafites. For xenoliths from kimberlites it has been shown by some authors (Boyd & Nixon, 1975; Meyer, 1977) that the sheared rocks are enriched in Fe, Ti and other elements. The degree of partial melting can be estimated by the decreasing amount of such mineral phases as clinopyroxene and spinel. The classification proposed by Pike & Schwarzman (1977) differs from those of other authors because they take into consideration also igneous and pyrometamorphic textures and propose criteria to distinguish them. The most important criteria for an igneous texture are: zonation of pyroxene, growth twins of pyroxene, and poikilitic relations. The pyrometamorphic texture is characterized by presence of glass formed by partial melting of rocks and the presence of quench minerals: olivine, clinopyroxene, plagioclase, and ore, as well as by spongy borders on pyroxene grains. The main types of metamorphic textures (porphyroclastic etc.) correspond to those of other authors.

Comparison of the textures of our rocks with those described above shows some similarities but also important differences. The textures of our rocks are porphyroclastic but relics of igneous textures are also present, as well as elements of a pyrometamorphic texture. Such a complex texture is the result of a multi-stage formation history of the rocks described in this paper. The relics of igneous textures are represented by the poikilitic relations of olivine and pyroxenes. Olivine included in pyroxenes differes from the olivine of porphyroclasts in the absence of any signs of deformation and the more pronounced degree of secondary alteration (serpentinization, carbonatization). Neoblasts of olivine show triple-point junctions and form typical mosaic aggregates. Thus, we have in our rocks three generations of olivine. Relics of an igneous texture are also grains of pyroxene with exsolution lamellae (of pyroxene), sometimes deformed along with the host mineral.

The pyrometamorphic (or 'decompressional'?) texture is represented by spongy border zones and grains mostly of clinopyroxene, but also of orthopyroxene and by desintegrated grains of clinopyroxene preserving a uniform optical orientation. In the classifications mentioned above, these types of processes (desintegration) are not mentioned but they are important and the 'aggregates' of clinopyroxene were mobile (vein-like form) and can have caused the yielding to a sort of plastic flow of the whole rock. The presence of quench minerals (Pl, Ti-Aug, iron-ore) is also an element of the pyrometamorphic texture.

Spinel-bearing lherzolites of Finkenberg with a porphyroclastic texture have several generations of major mineral phases (see above): 3 generations of olivine, 3 generations of orthopyroxene (exsolution lamellae partly recrystallized, porphyroclasts and neoblasts), 4 generations of clinopyroxene (exsolution lamellae, desintegrated grains, neoblasts, and quench-microlites), and 3 or 4 generations of spinel. Vermicular spinel is thought (Mercier & Nicolas, 1975) to be typical for protogranular texture but in our nodules this form is also typical for the porphyroclastic rocks. Other generations of spinel are: exsolution lamellae, small grains in association with decomposed clinopyroxene and neoblasts. These observations lead to the conclusion that the formation of our spinel-lherzolites was a multi-stage process: high-pressure and high-temperature crystallization, subsolidus equilibration, deformation, and recrystallization – all preceding the incorporation in basalts; the partial melting stage and the contact metamorphism are related to this incorporation.

Thus, textures of the nodules can give information about the formation history of these rocks. It is more difficult to quantify the temperature and the pressure of these processes. Various methods for estimating the temperature and pressure of equilibration for mineral assemblages in lherzolites have been proposed. These methods are based on the composition of minerals and partitioning of different elements over the major mineral phases. Two methods are usually utilized: the single pyroxene geothermometer (Herzberg & Chapman, 1976; Obata, 1976) and the two-pyroxene geothermometer (Wood & Banno, 1973; Mysen, 1976; Mori, 1977). Results obtained by different methods may differ (Table III). Donaldson (1978) estimates that such a difference can reach 120-300°C. It is difficult to take into consideration all factors which have influenced the rock formation process: PH_2O , presence of different chemical elements in pyroxenes, presence of Al-bearing phases, such as spinel, plagioclase in rocks with Al_2O_3 -rich pyroxenes, etc. The pressure is usually estimated from the Al_2O_3 -content in orthopyroxenes (Boyd & England, 1964; O'Hara, 1967; McGregor, 1969, 1974) but also by other methods, e.g. content of Cr_2O_3 and Fe in spinel (Oehm et al., 1983). It should be said that the pressure estimations are rather uncertain.

It is possible to judge the equilibrium conditions, in a very general way, on the basis of mineral assemblages typical for lherzolites. Thus, the usual four-phase association (OI + Opx + Cpx + Sp) indicates (Wallace, 1975) that these rocks were formed under pressures between 10 and 20 kb and at temperatures of 900-1000°C. Most rocks are probably formed at the low-pressure end of this interval. The assumption of relatively low-pressure conditions is supported by the lack of a stable olivine-plagioclase association in spinel-lherzolites (Ghent et al., 1980). The absence of garnet and plagioclase and the presence of spinel instead of plagioclase indicate that the P-T conditions must have been between 1000°C at 10 kb and 1100°C at 19-20 kb (Herzberg, 1978; Tracy, 1980; Brown et al., 1980). According to experimental data (Green & Hibberson, 1970; Green & Ringwood, 1970), plagioclase disappears from 'pyrolite' at 11-12 kb at 1200°C; garnet can be formed at 18 kb. The temperatures obtained by different authors (Table 3) are below the possible crystallization temperatures of major mineral phases from a melt. The average temperature varies between 900 and 1250°C, but 700-800°C is also common. These temperatures lie below the liquidus temperature of basalts (Frey & Green, 1974; Hutchison et al., 1975; Donaldson, 1979). It seems, therefore, that the mineralogical composition of rocks reflects only the subsolidus stage of their formation. Studies of rocks where mineral phases of all stages are preserved may possibly give a larger interval. In our case the temperatures can be estimated to have ranged from c. 1450 to 700°C, or even lower. It is difficult to estimate the pressure value.

Regarding the origin of spinel-lherzolite nodules, various hypotheses have been proposed. These rocks may represent: 1) cumulates precipitated by a basic magma, related or not to the host basalts; 2) primordial (fertile) mantle material from which basaltic melts may be produced by partial melting; or 3) depleted (barren) mantle material – residua left after extraction of basaltic magma.

The hypothesis of a cumulate (cognate) origin of spinel-bearing ultramafic nodules is not very popular nowadays. Only for pyroxene-rich rocks of this group and for pyroxenites such an origin is accepted (Carter, 1970; Hutchison et al., 1975; Wilkinson, 1976; Frey & Prinz, 1978). During the partial melting, clinopyroxene is the first phase to disappear. The high magnesium content of lherzolites and of their olivine and orthopyroxenes indicates that these rocks hardly can be cumulates. The limited compositional range of xenoliths found in mafic volcanic rocks of varying composition (from alkali-olivine-basalts to phonolites) is difficult to reconcile with an accumulation process from different melts giving cumulates of identical composition.

The differences in chemical composition of the same minerals from xenoliths and basalts show that lherzolite cannot be a product of accumulation of phenocrysts. Geochemical data, such as the ⁸⁷Sr/⁸⁶Sr ratio (Kleeman et al., 1969; Paul, 1971; Ridley & Dawson, 1975), values of σ^{18} O (Kyser et al., 1982), distribution of U, Th and K in lavas and xenoliths, as well as isotopic analyses of lead (Kleeman & Cooper, 1970), indicate that spinel-lherzolites and harzburgites cannot have cognate relationships to the host basalts and cannot have been formed as cumulates from any alkaline basaltic magma (Rodes & Dawson, 1975). On the basis of the bulk composition of nodules and major- and trace-element abundances, as well as on the basis of mineralogical and isotopic data, the ultramafic nodules of the lherzolite series are generally considered accidental xenoliths, representing upper mantle material. They can, however, be fragments of primordial mantle material, or residual material. Lherzolites rich in low-melting components must represent mantle material, having a composition close to the composition of the primitive mantle which is a potential source of basaltic magmas (Forbes & Kuno, 1965; White, 1966; Green & Ringwood, 1967; Kuno, 1969; Kuno & Aoki, 1970; Frey & Green, 1974). In contrast, ultramafic rocks rich in refractory elements (Mg, Cr, Ni, etc.) and minerals (olivine) can be considered residual material from which basaltic magma was extracted. Chemical, geochemical, isotopic, and experimental data permitted to establish some criteria for distinguishing these different types of mantle material (Kushiro, 1969; Trask, 1969; Kuno & Aoki, 1970; Mysen & Kushiro, 1977; Ghent et al., 1980; Dawson, 1981). On the basis of the MgO-content in rocks (>40% MgO: residues; Hutchison et al., 1975), or values of the ratio MgO/MgO + Σ FeO or Mg/Mg + Σ Fe (90-92 for residua), the type of mantle material can be definitely established. The variation in these values reflects the degree of partial melting.

Most refractory material is depleted in CaO, Fe_2O_3 , TiO_2 , and HREE. The continuous extraction of basaltic magmas from upper mantle material through the process of partial melting must lead to the formation of residues increasingly enriched in magnesium

Publication	area	method	T(°C)	P(kbar)	depth (km)	remarks
O'Hara, 196	7		1200-1300	10-14		
O'Hara, 197	1		940-1100 1070-1120			
Griffin, 1973	Norway	Mg/Fe partition in mineral phases	1200-1250	9-11	33	near Moho
Frey & Green, 1974	Australia	Al ₂ O ₃ -content (Opx) CaO-content (Opx) Cao-, FeO-, MgO- content (Cpx)	≤ 900 700-800 ≤ 1050			similar to temperature of high-grade metamorphism below the solidi of lherzolites and below the liquidi of natural basalts
Wallace, 1975	New Zealand	mineral assemblage absence of garnet comparison of Cpx composition with results of experiments	900-1300	10-19-22 below 20- 18	25	
Hutchinson et al., 1975	Massif Central	two-pyroxene geothermometer	950-1110			below the melting interval of basalts
Wilkinson, 1975a	Australia	absence of garnet CaO-content in Cpx	1100	<20-25		
Suwa et al., 1975	Kenya	Al ₂ O ₃ -content in Cpx	1090	16		
Dostal & Capedri, 1976	Sardinia		below 950			
Varne, 1977	Tasmania	Al ₂ O ₃ -content in spinel	900-1150		<60	
Herzberg, 1978		absence of garnet absence of plagio- clase	1100	19 10	40-50	
Frey & Prinz, 1978	Arizona	Al ₂ O ₃ -content in Opx two-pyroxene geothermometer absence of garnet absence of plagioclase	1200-1300 1100-1300	<25 < 9		

Table 3. P/T conditions of spinel-lherzolites.

Donaldson, 1978	Great Britain	comparison with the CMAS system rocks in synthetic systems	890-1340 890-1190 av.950 + 50	15 + 3	40-48	90-140 below the anhydrous soli- dus of peridotite and close to the water-saturated solidus, close to the spinel/garnet transition
Brown et al., 1980	Massif Central	Al ₂ O ₃ -content in Opx absence of garnet absence of plagioclase	900-1100 1100 1000	14 18 8	20-25	near the trans- ition Sp/Gr
Tracy, 1980	Tahiti	two-pyroxene geo- thermometer Al ₂ O ₃ -content in Opx subsolidus equilibra- tion absence of garnet	1100-1200 1025 750-850		50	
Ghent et al., 1980	Saudi Arabia	Wells method Mori method Herzberg method absence of plagioclase	950-1050 1110-1350 1160-1335 1000	10-20 20-24 10	70	
Girod et al., 1981	S. Algeria	single pyroxene geothermometer	1000-1100	20-25		
Hamad, 1982	Sudan		900-1130			
Larsen, 1982	W. Greenland		700-830			
Fujii & Scarfe, 1982	British Columbia	two-pyroxene geothermometer	920-980- 1040	10-18		
Oehm et al., 1983	Hessian Depression, Germany	partition Mg/Fe between coexisting phases Cr and Fe in spinel	950-1060 875	21 21	50-80	
Kuo & Essene, 1983	Saudi Arabia	Ca/Al in pyroxene Mg/Fe in pyroxene	900-1200 800-900			
Griffin et al., 1984	Australia	two-pyroxene geothermometer	950			<u>, , , , , , , , , , , , , , , , , , , </u>
Stolz, 1984	Australia	two-pyroxene geothermometer	870-890 subsolidus equilibra- tion		48-55	

and depleted in low-melting components (Carswell et al., 1979; Carswell, 1980; Dawson, 1981). High values of Mg/Fe and Cr/Al ratios are typical for the residues. This change in the chemical composition of rocks during the process of partial melting must also lead to a change in mineralogical composition of the residual rocks. The first phase to disappear in lherzolites or peridotites is clinopyroxene.

Experimental work (Mysen & Kushiro, 1977) has shown that the melting range of peridotite leads to a succession of mineral assemblages in residua. At a small degree of melting (<20%) the residuum is formed by olivine, orthopyroxene, clinopyroxene, \pm spinel – a four-phase lherzolite assemblage. At a higher degree of fusion (>20%) the residuum consists of olivine and orthopyroxene (harzburgite) and finally of olivine (dunite). Experiments of Takashi and Kushiro (1983) demonstrated that during the melting of lherzolite under a pressure of c. 25 kb clinopyroxene is absent from all rocks of this type. Because of the rarity of rocks containing less than 40% MgO it is admitted that most lherzolite inclusions represent residues of different degrees of melting rather than primordial mantle material.

Frey and Green (1974) have noted a discrepancy between the trends in abundance of LIL elements and in the abundance of the major and compatible trace element (Sc, Cr, Co, Ni, HREE) in lherzolites. This discrepancy – the enrichment in LIL elements of rocks with high values of the Mg/Mg + Σ Fe ratios (residua) – can be explained, according to Frey et al., (1978), by adopting a 'two component model'. Component A, forming the major volume of the lherzolites and determining the distribution of the major and compatible trace element abundances, is thought to have a residual character. Component B, enriched in incompatible elements, is a product of penetration in the depleted mantle of a fluid phase genetically not related to the component A. According to this model, the lherzolite nodules are formed as follows: partial melting and forming of the residual component A, penetration in this depleted mantle of a fluid enriched in incompatible elements, resulting in component B; cooling and recrystallization of both components under subsolidus high-pressure conditions and accidental incorporation in the ascending basaltic magma. It seems probable that ultramafic xenoliths of the lherzolite series can have different origins: the pyroxene-rich rocks (mostly clinopyroxenites) can be cumulates; lherzolites, harzburgites and dunites may represent residual material left after different degrees of partial melting. Lherzolites can also be fragments of the primordial mantle material.

As to the origin of the lherzolite nodules of Finkenberg with their multi-stage formation history, it seems most reasonable to consider them as accidental xenoliths not related to the host basalts. The composition of the nodules – presence of more than 10% of clinopyroxene but a relatively high-magnesian character of the olivine and orthopyroxene – indicates that the lherzolites are probably close to the composition of undepleted upper mantle material which had undergone a relatively low (< 20%) degree of melting.

Xenoliths of the wehrlite suite

The xenoliths of the wehrlite suite are usually regarded as cumulates cognate to the host lavas, or related to a previous melting episode of the mantle (Frechen, 1963; White, 1966; Aoki, 1968; Kuno, 1969; Best, 1970; Kutolin & Frolova, 1970; Dawson & Smith, 1977; Irving, 1980; Ghent et al., 1980). The distinct characteristics of the wehrlite suite are: the

great variety of rock types, the inhomogeneity in the distribution of minerals with a tendency to form monomineralic bands or lenses, the cumulus (poikilitic) texture and the presence of a Ti- and Al-rich augite as the major phase, but also of olivine, orthopyroxene, spinel, and often plagioclase and kaersutite.

The members of the wehrlite suite of this study are represented by wehrlites and clinopyroxenites (Table 4). Mostly they are phlogopite-bearing rocks. Apart from the abundance of phlogopite they are similar to those found as inclusions in alkali-basalts elsewhere in the world. However, they differ from the common type by some important characteristics, viz. the absence of orthopyroxene, spinel, plagioclase, and amphibole. Different types of clinopyroxenites are also included in this group. The phlogopite-bearing clinopyroxenites, formed by augite or salite, seem without doubt to belong to the wehrlite suite. But the coarse-granular clinopyroxenites, consisting of iron-rich green salite, and also those formed by an Fe-Na-rich augite and containing such accessory minerals as apatite and titanite and primary carbonate hardly may be regarded as related to the phlogopite-bearing wehrlites. They are included in this group only because they are also magmatic rocks like wehrlites and because no other generally accepted suite has yet been proposed for such rocks.

The presence of phlogopite in rocks of the wehrlite group of this study poses the question of the role of metasomatism in their formation. Petrological and chemical studies have shown the abundance of K, Ti and other incompatible elements in alkali-basalts relative to tholeiites. Therefore, it is generally accepted that the partial melting of upper mantle peridotites (lherzolites, or garnet-lherzolites) cannot give rise to an alkali-basaltic melt, and that the source area for these magmas in the upper mantle must be enriched in alkalis, TiO₂, H₂O, and other elements. Mineral phases containing these chemical elements could be amphibole and/or phlogopite. Experimental work has shown that these minerals are stable under mantle conditions and thus they can be important components involved in the partial melting of mantle material to produce alkali-basaltic melts.

The introduction of these components in the mantle rocks is thought to be related to the process of infiltration metasomatism precursory to the partial melting. For these reasons considerable attention has been accorded in the last 15-20 years to phlogopite-(and amphibole-)bearing xenoliths constantly associated with provinces of alkali-basaltic volcanism. Such xenoliths have been reported from different parts of the world (Frechen, 1948, 1963; Aoki & Kushiro, 1968; Wilshire & Trask, 1971; Richter, 1971; Lloyd & Bailey, 1975; Wilkinson, 1975; Becker, 1977; Lloyd, 1981). They have also been described from West Germany (Zirkel, 1903; Frechen; Aoki & Kushiro; Lloyd & Bailey; Becker). The samples described here come from Finkenberg (collection of Dr H. M. E. Schürmann).

PHLOGOPITE-BEARING ROCKS

Among the phlogopite-bearing xenoliths (11 samples) two types of rocks can be distinguished. The most common type is represented by wehrlites (7 samples); less abundant are clinopyroxenites (3 samples). One xenolith consists only of olivine and phlogopite. In most cases several generations of the major mineral phases can be noted. Relics of primary textures may be observed, but sometimes the rocks are strongly deformed and recrystallized. The degree of deformation as well as the intensity of the metasomatism are different in different rocks. Mineralogical and textural evidence of different kinds of rock-forming processes, from magmatic crystallization via processes of adjustment in response to new P-T conditions to late alteration processes, are described below.

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Table 2. Spinel-bearing ultramafic xenoliths.

The average modal composition of wehrlites is relatively constant: 45-50% olivine; 35-40% clinopyroxene, 15 (5-20)% phlogopite, and 2-3% iron-ore. The mineral distribution is often irregular and some specimens exhibit a sort of banding where olivine-rich rocks (75% olivine) alternate with clinopyroxenites (70% clinopyroxene). The quantity of phlogopite may vary strongly; this mineral may form in some parts of a rock numerous veinlets or veinlike aggregates, and in others only relatively rare separate flakes occur.

The textures of wehrlites are coarse-granular or porphyroclastic, but relics of poikilitic texture are usually observed. The coarse-grained wehrlites show weak deformation but in porphyroclastic rocks kink-bands, deformation lamellae and subgrains in olivine are widespread.

Coarse-grained wehrlites (RGM 319 396, 319 390, 319 038, 319 139)

Olivine — The forsterite (Fo₈₆; $2V\gamma = 85-87^{\circ}$) is often partly or completely replaced by secondary minerals: serpentine, talk, carbonate, and – rarely – iddingsite. Two types of olivine grains are present: 1) euhedral or rounded small (\emptyset 0.1-0.2 mm) grains poikilitically enclosed by large clinopyroxene grains, and 2) relatively large (\emptyset 1.0-3.0 mm) subhedral or euhedral crystals. Slight undulose extinction is common.

Clinopyroxene — It is a pale-green or colourless diopside (salite?) $(2V\gamma = 60-62^\circ; c \not\equiv \gamma = 39-42^\circ)$. Its anhedral grains fill up the space between olivine grains. The dimensions of the salite grains are up to 5.0 mm in diameter, but most frequently the diameter is 1.2 to 2.0 mm. The clinopyroxene is weakly deformed (undulose extinction). The salite contains abundant inclusions of various kinds. Swarms of tiny translucent and opaque inclusions intersect the grains in various directions. Among them melting cavities are probably also present. Sometimes zones of strain and of anomalous extinction are parallel to bands of inclusions. Along these zones fine veinlets of talc may have developed. Clinopyroxene poikilitically encloses small grains of olivine (mentioned above) and minor (\emptyset 0.05 mm) flakes of red-brown phlogopite with distinct pleochroism. This first generation of phlogopite (phlogopite I) is of particular interest because of its primary character (Pl. 6, fig. 1). It was probably one of the early near-liquidus phases precipitated from an alkali-rich and hydrous melt. Occasionally, the clinopyroxene encloses small grains of iron-ore. In some instances symplectitic intergrowths of clinopyroxene and ore have been observed.

A very typical process is the development of spongy rims on clinopyroxene and the 'desintegration' or breakdown of its grains. All successive stages of this process can be observed. At the beginning a border zone of low birefringence appears, showing a spongy texture with numerous cavities or inclusions (glass or plagioclase?). Such a spongy rim is developed not only along the grain boundaries but also around the poikilitic inclusions. Clinopyroxene in these zones is in optical continuity with the central parts of the grains. By further development of this process large grains became divided into small sectors, separated from each other but resting in optical continuity. The interstices between these sectors are filled by labradorite, iron-ore, rare flakes of phlogopite, as well as by secondary minerals such as serpentine, carbonate, chalcedony, and Fe-oxides. The large porous grains of clinopyroxene and the small sectors are often rimmed by a titanium-rich pink augite; this rim is particularly distinct when the interstices are filled by plagioclase. The pink rim is also present at contacts of the clinopyroxene grains with the host basalt.

Phlogopite -- It shows four generations of grains: phlogopite I is poikilitically included in pyroxene. Phlogopite II is the major mode of occurrence of mica in wehrlites. These are flakes of brown phlogopite (\emptyset 0.8-2.0 mm) with relatively weak pleochroism; they are often partly or completely replaced by secondary minerals. Phlogopite II forms veinlets or vein-like aggregates. It has equilibrium boundaries with olivine without any reaction. The relations between phlogopite and clinopyroxene can be various. Sometimes its flakes with smooth outlines occur in apparent textural equilibrium with pyroxene (Pl. 6, fig. 1) (RGM 319 390) and can be determined as 'primary metasomatic' (using the terminology of Harte, 1977). Mostly the textural relations between these two phases are rather active with irregularly embayed outlines of the grains; phlogopite penetrates along cleavage planes in clinopyroxene grains or fills cracks, clearly replacing clinopyroxene. This is a 'secondary metasomatic' phlogopite (Pl. 6, fig. 2). In some rocks symplectite-like intergrowths of phlogopite and ore have been observed. The phlogopite II is often replaced by late secondary phlogopite, serpentine, iron-ore, carbonate, and chalcedony. In three samples (RGM 319 390, 319 396, 319 394) the major replacing phase is apatite, which forms small and very numerous grains in lineair 'chains' along the cleavage planes of phlogopite II (Pl. 7, fig. 2). Phlogopite III is a late secondary, scurfy red-brown or deep-brown mica replacing phlogopite II flakes, mostly along cleavage planes along with other secondary minerals. Its secondary character is beyond doubt. Phlogopite IV is a needle-shaped dark, purple-brown, Ti-rich mineral forming thin marginal zones along with iron-ore on crystals of phlogopite II at contacts with the host basalts (Pl. 8, fig. 1).

Xenoliths of wehrlite as well as the host basalts are cut by veinlets of carbonate, serpentine and chalcedony often with Fe-oxides.

Porphyroclastic wehrlites (RGM 319 394, 319 025, 319 397)

The subdivision of wehrlites into textural types or groups is often rather arbitrary. Thus, the larger part of xenolith RGM 319 025 is a wehrlite with a poikilitic texture while the smaller one is a porphyroclastic dunite. The degree of deformation in this group of wehrlites is generally strong and processes of recrystallization, absent in coarse-granular types, are distinctly developed. But relics of a poikilitic texture can be seen in most rocks.

Olivine — The forsterite (Fo₈₅₋₈₆; $2V\gamma = 84-86^{\circ}$) shows three forms: 1) poikilitic inclusions (Ø 0.1-0.2 mm) in clinopyroxene, 2) porphyroclasts (Ø 2.0-2.5 mm, occasionally 1.5-4.0 mm), 3) grains (Ø 0.8-1.2 mm) forming the matrix together with clinopyroxene. The porphyroclasts have a rounded ellipsoidal shape and are often oriented in a sub-parallel way. Their contours are irregular, curvilinear often disturbed by minor cracks. The olivine is strongly deformed: kink-bands, deformation lamellae and subgrains are common. Secondary micro-inclusions form sometimes zones parallel to deformation lamellae. The secondary minerals replacing olivine are serpentine, carbonate and chalcedony. The grains of olivine in the matrix are mostly also deformed, but in some domains mosaic aggregates with 120°-triple-junctions are seen. This proves that the olivines are partly recrystallized.

Clinopyroxene — It is a colourless or light-green diopside (salite?) ($2V\gamma = 60-64^{\circ}$, $c \not = 39-43^{\circ}$). Two types of grains are present: 1) large (\emptyset 2.0-2.8 mm, occasionally up to 4.0 mm) oikocrysts with inclusions of olivine and phlogopite, rich in micro-inclusions; 2) grains (\emptyset 0.8-1.0 mm) interstitial to porphyroclasts of olivine. The large grains always have a spongy rim, but the process of breakdown of clinopyroxene is weakly developed. The main part of the pyroxenes forms grains filling up the space between porphyroclasts; these

grains have curved outlines and contain sometimes poikilitic inclusions of olivine. The clinopyroxenes are partly recrystallized showing a mosaic texture with 120°-triple-junctions.

Phlogopite — It is present in four generations, as described above. Phlogopite II replaces to a large extent the large grains of clinopyroxene (RGM 319 025, 319 394). It forms vein-like aggregates. Sometimes rounded small grains of olivine are present in a mass of phlogopite; probably they were included in clinopyroxene now replaced by mica. Only in one sample (RGM 319 397) equilibrium grain boundaries of phlogopite and pyroxene have been noticed. In two samples (RGM 319 394, 319 397) the flakes of phlogopite hug the contours of olivine porphyroclasts.

Clinopyroxenites

The clinopyroxenites are coarse-grained rocks formed of 60-80% clinopyroxene and 15-20% phlogopite; in one sample about 10% olivine is present. They are weakly deformed. We distinguish two subgroups: olivine-bearing clinopyroxenites and other clinopyroxenites.

Olivine-bearing clinopyroxenites

This rock (RGM 319 101) has a poikilitic texture.

Olivine — It is included in clinopyroxene and has a composition of Fo_{84.86} ($2V\gamma = 86-88^{\circ}$). Its grains (\emptyset 0.2-0.5 mm) are partly or completely replaced by serpentine, talc, carbonate, or – rarely – iddingsite.

Clinopyroxene — The salite $(2V\gamma = 60-62^\circ, c \measuredangle \gamma = 38-40^\circ)$ occurs in grains (\emptyset 2.5-4.0 mm) of irregular or isometrical form with a simple curved boundary, which is slightly deformed. Poikilitic inclusions of phlogopite I and olivine, as well as micro-inclusions of various kinds, are numerous.

Phlogopite — The phlogopite I is often replaced by phlogopite III, carbonate and ore. Spongy rims on clinopyroxene appear around the flakes of phlogopite I. Phlogopite II has a 'primary-metasomatic' character but forms also veinlike aggregates. Otherwise this rock is very similar to those described above.

Other clinopyroxenites (RGM 319 380, 319 088)

Clinopyroxene — This is a colourless salite (?) $(2V\gamma = 58-60^\circ, c \measuredangle \gamma = 40-43^\circ)$ which occurs in grains of varying size (\emptyset 0.2-1.0 mm) of irregular or isometrical shape. It is slightly deformed (undulose extinction) and contains numerous micro-inclusions and minor melting cavities as well as poikilitic inclusions of phlogopite I, similar to those described for the wehrlites. A particularly large amount of minor inclusions occurs in the clinopyroxene of RGM 319 088. They form bands or zones oriented in different directions and give to the whole grain a spongy appearance. Parts of grains with numerous inclusions have a deeper green colour than zones lacking them, but the optical orientation and the birefringence are the same. It should be noted that, in contrast to clinopyroxene described in wehrlites (spongy-textural mantle surrounding a clear core), the pyroxene of the clinopyroxenite (RGM 319 088) has a spongy core and clear marginal zones. The latter clear zones are located not only along boundaries of grains but also along micro-cracks cutting the grains. Their formation is probably due to a process of recrystallization apparently following the formation of the sponge texture. In other clinopyroxenites the formation of spongy marginal zones is as described in the werhlites, and so is the break-down of pyroxene.

Phlogopite — It is present in three generations: 1) phlogopite I included in pyroxene (relatively rare), 2) phlogopite II, and 3) late secondary phlogopite III. Phlogopite II (\emptyset 1.6-2.0 mm) can be determined as 'primary metasomatic' having textural equilibrium relations with pyroxene; processes of replacement are practically absent. Occasionally phlogopite III penetrates along cleavage planes in the pyroxenes. Vein-like aggregates of phlogopite III filling the spaces between the clinopyroxene grains are also present.

Olivine — In one sample (RGM 319 088) the formation of olivine along the grain boundaries of pyroxene and phlogopite can be observed (Pl. 7, fig. 1). Olivine forms 'chains' of small (\emptyset 0.3-0.5 mm) elongated grains oriented with their long axis perpendicular to the border of the nearest grain of phlogopite or pyroxene. The interstices between grains of olivine are filled by carbonate, chalcedony and Fe-oxides. The processes of breakdown of the pyroxene and the formation of olivine are different and probably unrelated processes, but in domains of 'desintegrated' clinopyroxene the formation of olivine is strongly developed and not only near the boundaries of pyroxene and phlogopite. Often olivine forms 'series' of subparallel elongate grains, the amount of newly formed olivine in these places being often very large.

The spaces between olivine, clinopyroxene and phlogopite grains are filled by plagioclase and carbonate with rare minor flakes of phlogopite and micro-grains of iron-ore. Parts of the desintegrated clinopyroxene as well as large grains of it are mantled by a pink-coloured titanium-rich augite. Near the contacts with the host basalt these processes are particularly well developed. The amount of olivine grains (which differ strongly from the phenocrysts of olivine in the basalts) is very large. Some areas consist of minor portions of desintegrated clinopyroxene with the spaces between them filled exclusively by the newly formed olivine. Near the contact with the basalt the differences between the salite and olivine of the xenolith and the augite and olivine of the basalt are evident.

Dunite

The rock (RGM 319 407) consists of 82% olivine, 15% phlogopite and 3-5% iron-ore, and has a coarse-granular texture.

Olivine — It is completely replaced by such secondary minerals as serpentine, carbonate, chalcedony, and opal.

Phlogopite — Phlogopite II (\emptyset 0.8-1.0 mm) occurs in flakes and aggregates, occupying the space between the grains of olivine. Reaction relationships between these minerals are absent. Phlogopite II is replaced by phlogopite III, chlorite, serpentine, apatite, and carbonate.

Discussion

The typical mineral assemblage of phlogopite-bearing xenoliths is: Cpx + Ol + Sp + opaques. Absence of such minerals as garnet, orthopyroxene, spinel, and amphibole should be stressed. An inhomogeneous distribution of minerals with a tendency to form mono-mineralic lenses is common. The texture of the rocks is mostly poikilitic; even in samples with a porphyroclastic texture relics of clinopyroxene oikocrysts are present. The texture of the clinopyroxenites is coarse-granular, but rare poikilitic inclusions of phlogopite in pyroxene may be observed. The degree of deformation is weak.

The metasomatic processes have considerably modified the primary texture since the amount of phlogopite in nodules is typically more than 10%. An important feature is the presence of several generations of major mineral phases related to different processes and different stages of the complex history of the formation of the nodules (Table 4).

Let us re-iterate briefly these generations described above in detail. Olivine occurs in the following forms: 1) poikilitic inclusions in clinopyroxene, 2) separate grains formed simultaneously with clinopyroxene, 3) minor crystals formed by a reaction process. In porphyroclastic rocks a fourth form must be added: 4) porphyroclasts and neoblasts.

Clinopyroxene occurs as 1) oikocrysts with inclusions of phlogopite, olivine and occasionally of iron-ore, 2) large grains without inclusions crystallized simultaneously with olivine, 3) neoblasts in porphyroclastic wehrlites, and 4) rims of titanium-rich augite on primary pyroxene. Probably the spongy, broken down clinopyroxene should be seen as yet an other separate generation because it is related to another process than the forms mentioned above; it has also a different composition.

Phlogopite is known in four generations: 1) poikilitically included in clinopyroxene, 2) metasomatic phlogopite; two principal types must be distinguished: a 'primary metasomatic' one which is in textural equilibrium with other minerals, and a 'secondary metasomatic' one forming veinlets and replacing clinopyroxene; 3) late secondary, replacing the two first generations of phlogopite, and 4) titanium-rich phlogopite, present only at contacts with the host basalt.

Probably several generations of opaques are also present.

Among the late secondary processes, the formation of a considerable amount of apatite, replacing phlogopite, is of particular interest, but it cannot yet be satisfactory explained.

It is evident from all data mentioned that different processes have led to the formation of the rocks in question. The principal processes appear to be: 1) crystallization from a melt, 2) metasomatism, 3) deformation and recrystallization, 4) re-adjustment to the changing P-T conditions, 5) late secondary processes, and 6) contact metamorphism. It is not always possible to establish the succession of all these processes and to give their full characteristics, but an attempt can be made. The presence of poikilitic textures shows that the major mineral phases crystallized from a melt. The order of crystallization is considered to be: $OI \rightarrow Phl + OI \rightarrow OI + Cpx$. The absence of phlogopite inclusions in olivine indicates that olivine was the first liquidus phase, followed by the simultaneous crystallization of phlogopite and olivine. Clinopyroxene was the next phase to crystallize. During a period of time these two minerals (OI + Cpx) must have crystallized simultaneously (in clinopyroxenites the order of crystallization was: Phl $\rightarrow Cpx$). Experimental work suggests the possibility of such an order of precipitation from the melt.

Many workers (Kushiro et al., 1967; Kushiro, 1969; Yoder & Kushiro, 1969; Flower, 1971; Modreski & Boettcher, 1972, 1973; Forbes & Flower, 1974; Bravo & O'Hara, 1975; Boettcher et al., 1975; Edgar et al., 1976) have shown that phlogopite is stable in a

relatively wide interval of temperatures and pressures. Experiments (Edgar et al., 1976, 1980; Arima & Edgar, 1983) have demonstrated that phlogopite is a liquidus phase in a variety of alkali-rich melts in the presence of H_2O and CO_2 and under pressures of c. 15-20 kb. Studies of the upper stability limit of phlogopite (Luth, 1967; Kushiro et al., 1967; Yoder & Kushiro, 1969) indicate that in highly alkaline liquids, in presence as well as in the absence of H₂O, phlogopite melts incongruently according to the reaction: Phl \rightleftharpoons Fo + liq or Phl + vapor \rightleftharpoons Fo + liq In the presence of enstatite and/or diopside (Modreski & Boettcher, 1972, 1973; Boettcher et al., 1975) the incongruent melting of phologopite (having reaction relationships with pyroxenes) takes place according to the equation: PhI + Di \geq Fo + liq (T = 1135-1300 °C, P = 10-30 kb), or PhI + En_{ss} + Di_{ss} \neq Fo + Pyrope + liq (T = 1350°C, P = 35 kb). Later, Wendlandt and Eggler (1980b) have confirmed that the reaction actually takes place: Phl + $liq_1 \gtrsim Fo + liq_2$ (in presence as well as in absence of the gas phase). Thus the results of experiments confirm the possibility of formation of olivine as a product of the reaction between the phases mentioned. This process can be held responsible for the formation of the secondary – reactional - olivine in the clinopyroxenites of our material.

Primary magmatic phlogopite has been reported from xenoliths in both kimberlites and alkali-basalts, but only rarely inclusions of phlogopite in clinopyroxene and/or olivine have been noted (Wilkinson, 1975; Harte et al., 1975; Dawson & Smith, 1977). Usually the conclusions about the magmatic character of phlogopite (often associated with amphibole) are based on its textural relationship to other minerals, which have been designated as 'equilibrium textural relations' with apparent lack of reaction (Aoki & Kushiro, 1968; Harte et al., 1975; Carswell, 1975; Lloyd, 1981). However, as it was stressed by Wilshire and Trask (1975), these criteria are not adequate and can often be interpreted otherwise. In any case a common type of phlogopite described in ultramafic nodules is a metamorphic one: 'primary metasomatic' (Harte et al.), 'secondary metasomatic' (Carswell), or simply 'metasomatic'. And the question discussed is whether or not the process of metasomatism predates the incorporation of xenoliths into the host magma. It seems to be the most reasonable proposal that in ultramafic xenoliths in basalts phlogopite may be formed both as a magmatic and as a metasomatic phase.

In our case there seems to be little doubt that a magmatic phlogopite, formed during the early stages of magmatic evolution, is present. On the other hand the major part of this mineral forms veinlets and shows clear reaction-relations to clinopyroxene with textural evidence of replacement. Therefore, it may be concluded that the evolution of a magmatic liquid rich in volatiles and incompatible elements took place with decreasing temperature and pressure and led to the formation of a hydrous late-stage liquid or fluid phase. The phlogopite had thus a long and complex formation history and different generations of this mineral must differ from each other also in their chemical composition, reflecting the evolution of the melt – (hydrous?) fluid – solution. It seems possible that the process of crystallization from a melt and the later metasomatism are related and that both predate the incorporation of the nodules in the basalt. The contact metamorphism of clinopyroxene and of all generations of phlogopite support this suggestion.

An important stage in the history of the nodules was the conversion of rocks in response to changing physical conditions related to the ascent of basalts. The decompression related to this process caused partial melting, decomposition and breakdown of clinopyroxene. The formation of spongy rims and the spongy habit of whole crystals are a manifestation of this event that probably took place during, or immediately after, the transport of xenoliths to the surface. These processes are described by nearly all investigators of xenoliths (Wilshire & Binns, 1961; White, 1966; Kleeman et al., 1969; Kutolin & Frolova, 1970; Carswell, 1975; Kozlowska-Koch, 1976; Donaldson, 1978). They are

interpreted as being due to decompression. In our case such an interpretation may also be accepted.

The optically visible decomposition of pyroxene with formation of tubes and cavities filled by plagioclase or glass and the absence of spinel as a product of such decomposition (always present in decomposed diopside in spinel-lherzolites) show that the clinopyroxene in wehrlites is probably not very rich in Al_2O_3 and, in contrast, contains a jadeite molecule.

The processes of deformation and recrystallization of wehrlites were relatively weak, they took place in the period of time between the formation of phlogopite II and the incorporation of the nodules in basalt (occasionally deformed flakes of phlogopite, contact metamorphism of recrystallized phases). It should also be stressed that there are no relations between the texture of the nodule and the presence of phlogopite – phlogopite is present in porphyroclastic ('sheared') as well as in coarse granular wehrlites (in contrast to the data of Nixon & Boyd, 1973).

The last important process is the interaction between the nodules and the host basalts. The petrographical observations show all signs of desequilibrium at the contact of these rocks. Chemical reactions taking place here were stimulated not only by the difference in composition of the rocks but also by the changing temperature and pressure. These conclusions are supported by the formation of rims of low-pressure titanium-rich augite, similar to the pyroxene of the basalt, on the clinopyroxene grains of the nodules, by the reaction of the phlogopite with lava to form magnetite and/or titanium-rich phlogopite (Phl IV) and by the presence of newly formed flakes of phlogopite in the basalt near the contact.

All data mentioned lead to the conclusion that wehrlites equilibrated under different P-T conditions than the host basalts. The differences in chemical composition - enrichment of wehrlites in K_2O , H_2O , P_2O_5 , and other elements, if compared with the basalts (lacking K-bearing mineral phases) – show that these xenoliths are not related to the enclosing lavas and therefore cannot represent high-pressure cumulates from the same magma. They appear to be related to another magmatic event prior to the formation of the basalt melt. The comparison of the wehrlites studied with wehrlites from other localities lead to the conclusion that they should be considered as members of the wehrlite, or black-pyroxene, suite. Comparison of the wehrlites of this study with phlogopite-bearing ultramafic xenoliths known from other areas is of particular interest. Such xenoliths are mainly recorded from kimberlites and related rocks. A special suite of nodules rich in phlogopite and amphibole - the so-called MARID suite - was proposed by Dawson (Dawson & Smith, 1977). However, the comparison with this type of xenoliths is rather difficult because of the specific composition of kimberlites (rich in K, and subsequently in phlogopite, rich in volatiles), which may influence the formation of phlogopite in xenoliths. The mineral assemblage typical for these xenoliths (Ol + Opx + Gr + Ru + Ho) differs strongly from those described here. Therefore our xenoliths should be compared with the phlogopite-bearing nodules from alkaline basalts. Such nodules have been recorded from different areas (Frechen, 1948, 1963; Aoki & Kushiro, 1968; Wilshire & Trask, 1971; Richter, 1971; Lloyd & Bailey, 1975; Wilkinson, 1975; Becker, 1977; Lloyd, 1981). The phlogopite-bearing rocks are mostly wehrlites and clinopyroxenites. The phlogopite is mostly thought to be a magmatic phase and the xenoliths cumulates cognate to the host alkali-basaltic magma. These conclusions are mostly based on the relationships between clinopyroxene and phlogopite: the absence of reaction phenomena between these minerals. Sometimes the cognate character of the nodules is proved with the aid of chemical data (Becker, 1977), such as the similarity of clinopyroxenes and olivines of the nodules with the phenocrysts of the host basalts or the composition of the clinopyroxene which must have been in equilibrium with an alkali-rich undersaturated magma, etc.

A clearly metasomatic phlogopite has been described by Wilshire and Trask (1971). Along with amphibole, it forms veins cutting the deformed and recrystallized peridotite (lherzolite, wehrlite) of the nodules. It has been stressed by Lloyd and Bailey (1975) that phlogopite may be a magmatic as well as a metasomatic phase. They also noted the absence of reaction phenomena between the hydrous and anhydrous phases, which is generally regarded as an important criterion for magmatic origin of phlogopite (and amphibole).

If compared with phlogopite-bearing wehrlite nodules from other areas of West Germany (Eifel, Siebengebirge, etc.), the xenoliths described here differ by the composition of the clinopyroxene (titanaugite in the Eifel), the absence of amphibole, and the texture (amphibole and phlogopite in the Eifel often poikilitically enclose clinopyroxene). To our knowledge several generations of phlogopite in one rock have not been reported from any area except for 'primary' and 'secondary' micas from xenoliths in kimberlites (Carswell, 1975).

It should be said that the term 'primary' or 'primary-textured' has been used by different authors in a different sense. Mostly a primary phlogopite is thought to be a product of magmatic crystallization (Frechen, Dawson, Aoki, Carswell, Lloyd, and Bailey) but also as 'rather a subsolidus product of equilibration' (Harte et al., 1975), or as a result of a 'prolonged annealing following incursion of a metasomatizing fluid' (Delaney et al., 1980). The term 'primary phlogopite' has often been used to designate 'mantle derived phlogopite' (Dawson et al., 1970; Boyd & Nixon, 1978), or mica formed before the inclusion of nodules in kimberlite (Boyd & Nixon, 1973) without any other genetic consideration. Another term was added: 'primary-metasomatic' (Harte et al., 1975) 'metasomatic' (Dawson et al., 1970; Boettcher et al., 1979), or a phlogopite of 'secondary origin but primary textural relation' (Harte et al., 1975) and finally 'a primary-looking phlogopite' (Boyd & Nixon, 1975). This type of phlogopite is thought to be a metasomatic phase, or to have crystallized from a 'fractionated fluid'. For secondary phlogopites terms used are: secondary-metasomatic, secondary and late-secondary.

Without doubt it is difficult in many cases to draw conclusions concerning the origin of phlogopite. The nodules (especially those from kimberlites but also from alkali-basalts) were affected by a complex sequence of processes, such as metasomatism, deformation and recrystallization. The primary textures and the primary character of the minerals were strongly modified. So, for example, as noted by Carswell (1975), a clearly secondary phlogopite from kelyphitic rims around garnet was recrystallized to coarse, chemical uniform grains practically undistinguishable from a 'primary-textural' one. The chemical data are sometimes also insufficient for distinguishing primary from secondary phlogopite (Delaney et al., 1980). The petrographic observations show that the formation of phlogopite was a long and multi-stage process. During this process the temperature and pressure as well as the character of the mobile phase must have changed. Generalizing it can be said that the early phases were magmatic phases of liquid-crystal equilibrium while the late ones were related to a hydrous phase.

The four generations of phlogopite, described for the nodules from Finkenberg, can be determined as follows: Phl I is a primary or magmatic phlogopite, a product of crystallization from a melt. Criteria for this conclusion are: inclusions of phlogopite in other magmatic minerals, the textural equilibrium between phlogopite and other magmatic minerals: no veinlets of phlogopite, no reaction relationships with other minerals, no replacement phenomena. Phl II is a metasomatic phlogopite precipitated by a volatile-rich phase: a residual melt or fluid. Criteria: veinlets and veinlike aggregates of phlogopite, replacement phenomena. Phl III is a secondary phase forming rims, replacing earlier phlogopite along with serpentine, chlorite, carbonate, etc. Phl IV is a product of contact metamorphism. It seems reasonable to utilize the term 'primary' only for the magmatic phlogopite. In our case phlogopite is one of the major phases related to the formation of the whole rock and to the evolution of the mobile phase. For the cases where there are no genetic relationships between phlogopite and the rock, phlogopite should be called 'metasomatic'.

CLINOPYROXENITES

Xenoliths of clinopyroxenite from basalts have been reported from many areas of the world. They are not so common as the nodules of lherzolite or wehrlite. Clinopyroxenites occur as individual xenoliths as well as in the form of complex nodules consisting of both olivine-rich and pyroxene-rich ultramafic rocks. Pyroxenites of the lherzolite suite are rather scarce. In complex nodules they form thin bands in olivine-rich rocks. These pyroxenites consist of Cr-diopside with a minor amount of enstatite and may also contain olivine, spinel and opaques. Websterites are more often found. Typical for these rocks are their metamorphic texture and abundant evidence of exsolution processes. In spite of the absence of distinct magmatic textures it is widely accepted that pyroxenites have been formed by crystallization from a basaltic magma and they are regarded as cumulates (Hutchison & Gass, 1971; Wilshire & Pike, 1975). They are thought not to be related to the host basalts. More common are xenoliths of clinopyroxenite in the wehrlite suite. They have been reported by many investigators (Wilshire & Binns, 1961; White, 1966; Aoki & Kushiro, 1968; Kuno & Aoki, 1970; Ellis, 1976; Kozlowska-Koch, 1976; Conrad & Kay, 1984; Griffin et al., 1984). The main constituent of these rocks is clinopyroxene, usually an Al- and Ti-bearing augite or salite. Chemical analyses of the pyroxene show a relatively high content of Al₂O₃ and of Ca-Tschermak's component which induced Wilkinson to call it 'subcalcic pyroxene'. Often the presence of the jadeite-molecule has been established. Besides clinopyroxene, olivine, spinel, iron-ore, and in many cases also amphibole (pargasite or kaersutite) and/or phlogopite are present. The texture of clinopyroxenites is poikilitic or hypidiomorphic.

It is widely accepted that clinopyroxenites are cumulates which can be cognate with the host basalts (Best, 1970; Irving, 1974; Becker, 1977; Ghent et al., 1980), or are related to another magmatic event (Wilshire & Shervais, 1975; Wilkinson, 1975). A very special group is that of the clinopyroxenites with green clinopyroxene, which is enriched in Fe and often contains jadeite or acmite molecules. Clinopyroxenites of this type, as well as megacrysts of green clinopyroxene in alkali-basalts, have been described from many areas (Frechen, 1948, 1963; Babkine et al., 1968; Frisch & Schmincke, 1969; Borley et al., 1971; Huckenholz, 1973; Brooks & Rucklidge, 1973; Wass, 1979, 1980; Barton & van Bergen, 1981). The specific composition of these rocks and the presence of volatile-rich mineral phases, such as amphibole, mica, apatite, carbonate, etc. may indicate that metasomatism processes were of great importance for the formation of this type of clinopyroxenite. Their relationships to the wehrlite suite are not clear. In this section two groups of clinopyroxenites, differing from each other by the composition of clinopyroxene and accessory minerals, are described. The first one includes rocks consisting of Ti-rich augite and opaques. The clinopyroxene of the second group is a Fe- and Na-rich green augite or salite and the accessories are opaques, apatite and titanite. It must be noted that the mineral assemblages of these rocks differ from those of the clinopyroxenites of both the Cr-diopside and Al-augite xenolith-suites. Usually clinopyroxenites contain one or more of such minerals as olivine, orthopyroxene, amphibole, spinel, phlogopite, and occasionally garnet. However, in the clinopyroxenites studied these minerals are totally absent.

Group I

The texture of these rocks (RGM 319 350, 319 352, 319 368, 319 009) is coarse-granular and is characterized by euhedral grain shapes modified by mutual impingement during growth. No deformation processes are noted. The rock consists of 93-98% (in one case 85%) augite and 2-7% (in one sample 15%) iron-ore (probably ilmenite). Sample RGM 319 009 is composed of a greenish brown clinopyroxene and the rock is probably recrystallized, but the accessory minerals and other particularities of the rock allow to include this sample in Group I.

Clinopyroxene — It is a titanium-rich augite $(2V\gamma = 60-62^\circ; r>v; c \measuredangle \gamma = 43-44^\circ)$. In most samples it has a pale-brown or pale purplish brown colour. A complex colour-zoning can be observed: patches of green clinopyroxene occur in different parts of large grains and are surrounded by strongly predominating purplish brown augite. The green pyroxene differs from the green pyroxene of apatite-bearing pyroxenites by its colour, absence of pleochroism and degree of dispersion. It is probably the Fe-rich salite described also from other areas (e.g. Frechen, 1948, 1963). It should be noted that a similar zoning occurs in phenocrysts of the host basalts: a pale green core is surrounded by a narrow rim of purplish titanaugite. The zoning of clinopyroxene in xenoliths is often very complex and irregular, and can be seen under parallel as well as under crossed polarizers. The dispersion of the optic angle is strong.

The augite forms grains of different sizes (\emptyset 1.2-1.5 mm) but large grains usually predominate. A typical feature of clinopyroxene is the abundance of cavities and inclusions of various kinds. Small (\emptyset 0.1 mm) grains of iron-ore are occasionally included in pyroxene. Micro-inclusions, forming zones and bands, crossing grains of augite in various directions and intersecting one another, are very numerous. The nature of these inclusions can not always be determined, but it is certain that in these zones melting cavities are also present. Sometimes vermicular micro-cavities form separate zones. Another kind are small (\emptyset 0.05-0.1 mm) isolated cavities of a circular or irregular form filled by carbonate, which seems to be primary. Such cavities may be connected by thin cracks, or are located along cracks and form locally a network of intersecting zones. These cavities are filled by carbonate, iron-ore and Fe-oxides; small flakes of brown mica are occasionally present. Larger cavities (up to 1.0 mm) are filled by carbonate and iron-ore and a sort of zonation can be seen: grains of ore are located along border zones of host pyroxene, and carbonate fills the central part of the cavity.

The augite contains also inclusions of trapped liquid. The size and the form of this sort of inclusions is variable. Their form may be irregular, circular, or vein-like. The size varies from \emptyset 0.2 up to 4 mm. The trapped liquid is now usually represented by an aggregate of microlites of labradorite and minor grains of iron-ore. Rare flakes of mica and thin needles of pale-green amphibole have been observed. Serpentine and carbonate are nearly ubiquitous. Here also the same zonal distribution of minerals can be seen: iron-ore forms a rim around the host pyroxene which has a narrow distinct purple rim, while the centre is occupied by plagioclase and carbonate. The augite grains develop spongy border zones with numerous small vermicular cavities filled by plagioclase, carbonate and iron-ore. These zones can cross pyroxene grains. Some grains are completely converted into a spongy texture.

The augite has undergone a very particular recrystallization; the coincidence of spongy zones with domains of recrystallization indicates that the processes are spatially related. In different parts of the augite grains small (\emptyset 0.1-0.4 mm) patches or sections of the same mineral with a clearly geometrical form (rhombs) can be seen. The boundaries

of these sections with the host augite pass along micro-cracks and are underlined by small linear grains of opaques located along them. These sections may have the same colour as the host pyroxene and the same optical orientation but they usually show a higher birefringence. Such sections are absent in the green-coloured parts of clinopyroxene grains. The ubiquitous association of sections with cavities filled by carbonate and with cracks, as well as their linear form (rhombs form often a linear zone), indicate that the process of recrystallization took place along zones of 'weakness'. The concentration of iron-ore and often also of mica in these places shows that augite probably lost Fe and Ti during recrystallization.

It should be noted that linear zones of micro-inclusions, as present in the augite, can never be followed in the sections of recrystallized pyroxene and are interrupted near the boundaries of these sections. This fact confirms the suggestion that we deal here with a process of late recrystallization.

All particularities of the augite: the abundance of inclusions and cavities, patched, complex zoning, presence of spongy zones and of 'sections of recrystallization', give it a mosaic habit.

At the contact of the nodules with the host basalt augite is overgrown by a purplish salite with euhedral outlines. Sometimes a multiple zoning can be observed: the brownish salite of the nodule has a narrow rim of pale-green pyroxene, mantled by a low-pressure titanaugite.

The clinopyroxene contains a large amount of carbonate which fills cavities and cracks and forms veinlets crossing the whole rock. There are several generations of carbonate: primary as well as secondary.

Ore — Without doubt there are also different generations of ore minerals: 1) minor grains included in augite, 2) xenomorphic grains (\emptyset 0.5 mm) filling interstices between pyroxene grains, 3) minor grains precipitated from the trapped liquid, 4) iron-ore formed during the process of recrystallization of clinopyroxene, and 5) ore minerals filling cracks in augite grains.

The absence of deformation of pyroxene must be stressed. Occasionally a slight undulose extinction can be observed.

The petrographic study reveals that the clinopyroxenites of Group I are magmatic rocks precipitated from a basaltic melt. This conclusion is based on the igneous texture of the rock, with inclusions of primary ore minerals in augite, the abundance of cavities (in pyroxene) filled by plagioclase and primary carbonate, and on the presence of trapped liquid in clinopyroxene. The post-cumulus crystallization of trapped liquid can be clearly observed (low-pressure crystalline phases - plagioclase and titanaugite). The clinopyroxene shows a complex growth history. The crystallization started probably with an Fe-rich salite and was followed by the formation of a titanium-rich augite composing the predominant part of the magmatic pyroxene. The clinopyroxene of the nodules is compositionally similar to the phenocrysts of the basalt and they may have precipitated from similar melts. The absence of olivine from the xenoliths and its presence in the basalts indicates that the crystallization conditions were different. This difference concerns not only pressure and temperature but also the presence of volatiles (H₂O in the first place). It is known (Green, 1973) that H₂O can influence the presence or absence of olivine as a liquidus phase. In the presence of H2O and under pressures of about 20 kb clinopyroxene may be the sole liquidus phase. An accumulation process can lead to the formation of clinopyroxenites.

The present complex structure of the clinopyroxene is a result of both a crystallization under unstable physical conditions (changing pressure, temperature, PH_2O , etc.) and later processes of re-equilibration in response to changing P-T conditions, related to the ascent of basaltic magma, containing nodules, to the surface.

All these processes: partial melting of clinopyroxene (spongy rims), its recrystallization under relatively high temperature leading to the formation of a new pyroxene (with higher birefringence), and finally the formation of the low-pressure titanaugite rims, changed the primary habit of the magmatic pyroxene. The absence of equilibrium between nodules and the host basalts (multiple rims on pyroxene at the contact) confirms the suggestion that these rocks were formed under different P-T conditions. On the other hand, the similarity in composition of the pyroxenes (phenocrysts in basalts and pyroxenes in nodules) suggests that a genetic relation between them is rather likely. The basaltic magma may have lost the volatiles by its ascent to the surface and the liquidus phases under low-pressure conditions were consequently different.

Group II

The rocks (RGM 318 929, 319 366, 319 967) are formed by a green clinopyroxene (90-96%) and contain as accessories; iron-ore (RGM 319 367), iron-ore + apatite (RGM 318 929), or iron-ore + apatite + titanite (RGM 319 366). The texture of the rocks is coarse-granular. Clinopyroxene may be present as megacrysts with some smaller grains or as an aggregate of grains of different sizes (\emptyset 2.0-8.0 mm).

Clinopyroxene — It is an Fe- and Na-rich augite (or salite) $2V\gamma = 56-60^\circ$; r>v; $c \not\equiv 42-44^\circ$). It has a bluish green colour and a distinct pleochroism: $\alpha =$ greenish, $\beta =$ green, $\gamma =$ green-bluish. A complex patchy zoning is typical. The clinopyroxene is rich in inclusions and cavities. The inclusions are represented by ore minerals, apatite and titanite. Grains of ore have an irregular or circular form and are of various sizes ($\emptyset \ 0.03-2.5 \text{ mm}$). Euhedral crystals of apatite or groups of such crystals are common as inclusions in pyroxene. Three grains of titanite, enclosed by pyroxene, have been noticed. As in all pyroxenites, the most abundant are micro-inclusions, forming linear zones cutting the host grains in different directions.

Extremely numerous are cavities of various sizes and forms. They are mostly filled by carbonate. The most abundant are isolated circular cavities filled only by primary (?) carbonate. Others are filled by carbonate and ore, or carbonate + ore + plagioclase. Cavities related to the process of partial melting and recrystallization of pyroxene (see below) are also filled by carbonate. Inclusions of trapped liquid are numerous. Very often grains of clinopyroxene have a complex structure: zoning, spongy rims, inclusions, multiple overgrowths, and recrystallization zones are rather ubiquitous.

In clinopyroxenites with ore as accessory mineral a rim of purplish augite mantling the green augite can be seen. This rim is rich in cavities filled by carbonate and has a higher birefringence than the central part of the grain lacking such cavities. These rims are located along the borders of clinopyroxene grains but are especially strongly developed at the contact with the trapped liquid. The main particularities of the rims are: the abundance of micro-cavities, the presence along their boundaries of grains of ore and carbonate, outlining this boundary, and the purplish colour. Usually (in lherzolites, wehrlites and other clinopyroxenites) a pink rim overgrows the spongy rim and is free of cavities. This observation leads to the conclusion that the composition of the purplish clinopyroxene in the rim is different from the low-pressure titanaugite of basalts, because the former has undergone the process of decompression and partial melting (spongy rims). Especially complex is the structure of the clinopyroxenes in apatite-bearing clinopyroxenites. Two types or parts (sections, patches) of pyroxene grains can be distinguished here: 1) parts without cavities but rich in micro-inclusions, and 2) parts rich in cavities but lacking micro-inclusions; linear zones of micro-inclusions disappear near the cavity-rich parts. In most cases sections rich in cavities form a sort of rim mantling the grains, but they can also be observed in the central parts of the grains, often around inclusions of ore. The relative volume of the parts rich in cavities is highly variable; sometimes the rim forms only 1/10 of the diameter of the whole grain, in other instances more than 3/4. The whole grain preserves usually a uniform optical orientation, but the colour of the rims is more intense and the birefringence higher.

Very often a rhythmical zoning can be observed; especially in rims rich in cavities. This zoning is parallel to the contours of the rims and can be distinctly observed under parallel and crossed polarizers. The multiple overgrowth probably reflects a complex process of crystallization from a melt of changing chemistry (Pl. 8, fig. 2). Along contacts against trapped liquid as well as against the host basalt, clinopyroxene is mantled by a thin rim of pink titanium-rich augite. Sometimes the cavity-rich pyroxene is first bordered by a thin pale-green rim followed by a pink one.

Apatite — The amount of apatite in clinopyroxenites may reach 8-10%. Its crystals are usually euhedral and the length of the needle-like grains may reach 4.5 mm. Individual grains or groups of grains can be included in the central parts as well as in the cavity-rich rims of clinopyroxene crystals. Sometimes minute apatite grains form a sort of 'chain' cutting the pyroxene; in other instances they are localised along cracks and form veinlets along with ore and carbonate. Seldom apatite grains can be observed in trapped liquid in pyroxene.

The apatite contains fine indeterminable opaque minerals concentrated in two planes: parallel and perpendicular to the long axis of the crystals. Sometimes these inclusions are concentrated in the border parts of the grains, forming opaque rims; they can also cover the whole surface of grains. Numerous inclusions of liquid or gas (CO_2 ?) can be observed.

Titanite — Grains of titanite are present only in one sample (RGM 319 366), they are enclosed in pyroxene which develops a bright pink-coloured rim around them. One titanite grain contains a small crystal of apatite as an inclusion.

Ore minerals — They are present in several generations: 1) inclusions (\emptyset 0.5-1.0 mm) in pyroxene; 2) grains (\emptyset 1.0-4.0 mm) filling interstices between augite grains; 3) micrograins in trapped liquid, and 4) as veinlets cutting the clinopyroxenes.

Inclusions of trapped liquid are present in large grains of clinopyroxene, or fill spaces between them. They have an irregular form and various sizes (\emptyset 0.5-1.0 mm). Trapped melt is now represented by an aggregate of labradorite, iron-ore, carbonate, rare clinopyroxene, and rare flakes of mica; occasionally chalcedony and zeolites are also present. Labradorite can form relatively large (\emptyset 0.5 mm) grains of irregular form but occurs mostly in microlites. Magnetite grains form a rim along the borders of the host pyroxene which in that case has a pink rim. In instances where pyroxene grains occur they are euhedral and have a distinct hour-glass structure. Plagioclase and carbonate compose 80-90% of the volume of the trapped liquid. Zeolites and carbonate form veinlets cutting the nodules and are often located at the contact with the enclosing lava.

The host basalts contain phenocrysts of olivine and titanaugite occasionally with cores of green salite. Sometimes the basalts contain circular forms (amygdales?) resembling leucite crystals replaced by chalcedony and occasionally by zeolites.

Monomineralic apatite rocks

These rocks (RGM 319 164, 319 400, 319 297) form veins or lenses in apatite-bearing clinopyroxenites. At their contact, crystals of zoned green clinopyroxene enclosing euhedral crystals of apatite are cut by monomineralic aggregates of apatite. No visible contact reactions have been noticed. Xenocrysts of clinopyroxene are present in apatite. They have an irregular form and are often resorbed.

Apatite — The rocks are formed by large (\emptyset 2.5-3.0 mm) crystals of apatite, surrounded by a mass of small grains (\emptyset 0.3-0.8 mm) of the same mineral. Morphologically this texture is similar to a porphyroclastic texture. The relative amount of large and small grains varies from 90 to 15% of large grains passing through a ratio of about 1:1. The apatite seems to be a carbonate-bearing apatite with a relatively high (0.010) birefringence and is occasionally replaced by carbonate.

The large grains are elongated parallel to the axis \underline{c} and show a subparallel orientation. They are deformed, and undulose extinction, cracks and subgrains are common. The apatites contain inclusions of various kinds. Very numerous needle-like micro-inclusions of iron-ore give a black or blue colour to the large grains. These inclusions are parallel to the axis \underline{c} of the apatite. The amount of iron-ore needles can vary: sometimes they cover the whole grain but mostly they are concentrated in the central parts of the crystals. Black border zones, extremely rich in inclusions, can occasionally be observed. Mostly they occur at the contacts of apatite rocks with clinopyroxenites and also with host basalts. The iron-ore needles are often concentrated along microcracks and boundaries of subgrains, and then the parts of the apatite crystals near these cracks are free from inclusions. Such relations seem to indicate that primary inclusions are later redistributed whithin the grains. Iron-ore inclusions are absent in places of partial melting (or dissolving) of apatite and in zones where this mineral is party replaced by carbonate.

Another type of primary inclusions, oriented also in the direction of the longer axis of apatite grains, are micro-bars or rods of translucent inclusions of an unknown nature. Occasionally these inclusions are localized along cracks.

The apatite contains also abundant liquid and gas inclusions. They have various forms: isometrical, elongate, irregular, and vermicular. Two principal types can be distinguished: 1) 'capsuled' inclusions with a clearly visible outline which makes them similar to minerals with a high refractory index, and 2) 'drops' without such a clear outline, showing the same 'refractory index' as the host apatite. The first type is much more widespread. The distribution of both types of inclusions is rather irregular; they sometimes form bands parallel to the long axis of the apatite grains, but mostly they are related to micro-cracks and boundaries between subgrains. They can be located in cracks, or form bands near the cracks.

In large apatite grains, areas of low refractory index and low birefringence (relative to apatite) can be seen. These areas often have a very complex form; they are sometimes related to linear cracks, and have a vein-like form. It looks as if areas of melted apatite are free from iron-ore inclusions. It is possible that isotropic parts of these areas are glass. The large grains of apatite with a complex internal structure (described above) are surrounded by a sort of granoblastic aggregate of small grains of apatite, where occasionally triple point junctions of c. 120° can be observed. These grains contain usually less iron-ore inclusions than the large grains, and the inclusions are often concentrated only along cracks or between apatite grains. These grains show no signs of deformation.

Because of small amounts of iron-ore inclusions small grains of apatite are often colourless or pale-blue. Liquid inclusions and melted areas occur also in the aggregates of small grains. Here the process of carbonatization of apatite is more developed than in large grains.

The whole rock is often cut by veinlets of carbonate and zeolite.

The apatite-bearing clinopyroxenites are magmatic rocks with a magmatic texture. Textural relations between the minerals suggest that the order of crystallization is as follows: Ap + Ore (?) \rightarrow Ap + Sph \rightarrow Cpx. During crystallization, portions of melts (trapped liquid) with Ap or without crystalline phases were included in clinopyroxene grains. The specific mineral assemblage (green Cpx + Ap ± Sph) and the relations to the monomineralic apatite rocks indicate that the clinopyroxenites have a pronounced alkaline (sodic) character; therefore it may be thought that the green clinopyroxene contains a relatively high abundance of the acmite molecule. The abundance of carbonate present, as primary inclusions in pyroxene, as liquid and gaseous inclusions in apatite, and as filling of numerous cavities of all kinds, is a supplementary argument for the suggestion of the particular character of the parental melt rich in alkalies (Na), Fe, Ti, Cl, P, and CO₂.

The complex internal structure of the clinopyroxene crystals: their patchy and rhythmic zonation, the abundance of inclusions and cavities, as well as the evidence of re-equilibration processes indicate that the formation of these rocks was a long and complex process. It began with the precipitation from a melt of changing chemical compostion influenced probably by changing pressure, temperature, PH₂O and PCO₂. After the incorporation in basaltic magma ascending to the surface, processes of decompression have led to partial melting of the clinopyroxenes (spongy rims) and their recrystallization (two types of sections in pyroxene grains). These processes were followed (or accompanied) by the crystallization of the trapped liquid to a low-pressure mineral association (plagioclase, titanaugite) and by the formation of multiple zoning in contact with host lavas. The absence of equilibrium between the nodules and host basalts is evidence for the difference in conditions under which these rocks were formed. The difference in mineral phases present (olivine and titanaugite in basalts; green acmite-rich clinopyroxene in nodules, as well as the absence of olivine and the presence of apatite and titanite) and in the composition of the whole rocks show that there is no reason to suppose any genetic relation between them. The parental magma of the clinopyroxenites may have been related to an earlier magmatic event.

A green pyroxene in xenoliths has been recorded by many authors from different areas of basaltic volcanism (Frechen, 1948, 1963; Aoki & Kushiro, 1968; Frisch & Schminke, 1969; Borley et al., 1971; Wilkinson, 1975; Brooks & Printzlau, 1978; Wass, 1979; Wass et al., 1980; Barton & van Bergen, 1981). Chemical analyses of this mineral show a high content of the acmite molecule 10% (Babkine et al., 1968), 15% (Aoki & Kushiro, 1968), 30% (Griffin & Taylor, 1975) as well as of the Ca-Tschermak's and Ferri-Ca-component. This clinopyroxene is considered as acmite-rich salite (or augite), or as 'salite ferri-sodic' (Babkine, 1968). In many examples, reported from different localities, the green clinopyroxene is associated with amphibole (pargasite or kaersutite). Wass et al. (1980) have observed a green clinopyroxene replacing titanaugite. Clinopyroxenites with green pyroxene often contain phlogopite, apatite and spinel. A special type

of xenoliths, containing usually only a small amount of green clinopyroxene and formed by amphibole + apatite, has been described by some investigators (Wass, 1979; Watson, 1980; Wass et al., 1980). These authors consider this type of xenoliths as crystallizate from a residual fluid of a kimberlitic melt and as rocks similar to the MARID suite of xenoliths of Dawson & Smith (1977), interpreted as cumulates. These xenoliths differ strongly from the clinopyroxenites of this study. The xenoliths of clinopyroxenites with a green pyroxene are often associated with basaltic lavas of potassic affinity, containing phenocrysts with a core of green pyroxene. Such an association has been noted by many authors (Babkine et al., 1968; Frisch & Schminke, 1969; Scott, 1976; Thompson, 1977). It is generally accepted that the composition of all these pyroxenes is similar, or identical, and therefore a genetic relationship exists between the clinopyroxenites and the host basalts. But it has also been shown (Huckenholz, 1973; Larsen, 1977; Brooks & Printzlau, 1978; Barton & van Bergen, 1981) that in some cases the clinopyroxene can not have crystallized from the melt now represented by the host basalt, and hence any genetic relationship must be excluded. Some authors (Aoki & Kushiro, Wilkinson, Wass) consider the green pyroxene as a product of crystallization from the host basalt under higher pressure (i.e. as a member of a high-pressure trend of the Al-augite series). Others (Scott; Barton & van Bergen) exclude a genetic relationship between the green pyroxene and the basalt and consider its megacrysts, as well as the xenocrysts, as fragments of desaggregated veins or metasomatic lenses in the wall rocks, or as fragments of a metasomatized mantle. The points of view on the origin of the green pyroxene-bearing clinopyroxenites, thus can be summarised as follows: 1) crystallization from the host-basalt magma under higher pressure: 2) high-pressure crystallization from another magma; 3) crystallization from a kimberlite/carbonatite melt/fluid (close relation to the MARID suite); 4) result of a process of mixing of different magmas (Brooks & Printzlau, 1978); 5) fragments of the metasomatized mantle.

Thus, the origin of these clinopyroxenites is not yet satisfactory understood. Clinopyroxenites of this type from our collection differ from most xenoliths of this kind described elsewhere. They contain no hydrous minerals and no spinel. The absence of replacement phenomena, observed in other cases, shows that metasomatism has not played an important role in the formation of these rocks. The texture of the clinopyroxenites indicates that they have been formed as products of crystallization of a melt enriched in such elements as Fe, Na, Ti, P, Cl, and also CO_2 and H_2O . Such a melt was probably a product of partial melting of a strongly metasomatized mantle.

CONCLUSIONS

All ultramafic nodules described in this chapter are magmatic rocks with an igneous texture, or with relics of such a texture. The later processes of metasomatism and recrystallization have modified in different degrees the primary textures, but in most cases the formation history of the rock can be reconstructed. This reconstruction shows that it was a complex and multi-stage process. Magmatic crystallization, reactions between mineral phases, recrystallization, replacement phenomena, partial melting, and contact metamorphism were the most important events, which have given the rocks their present habit.

The phlogopite-bearing rocks belong to the wehrlite series and exhibit features typical for this xenolith suite: a magmatic texture, inhomogenous distribution of minerals, absence of orthopyroxene, and presence of augite, indicating a relatively weak deformation. A particular characteristic of these rocks is the strong development of metasomatism which led to the formation of phlogopite as a rock-forming mineral and as a metasomatic phase. Wehrlites lacking phlogopite are absent.

The second type of xenoliths – the alkaline (sodic?) clinopyroxenites with apatite, titanite, carbonate, and lenses of monomineralic apatite rocks – are included in this suite conditionally (see above). They cannot be regarded as members of the suite because of their very specific mineralogical and chemical composition. It is also difficult to consider them as belonging to the amphibole/apatite suite of Wass (1979, 1980) or to the MARID suite (Dawson & Smith, 1977). The major mineral phases typical for these suites (amphibole, mica, rutile, etc.), and the replacement phenomena developed there, are totally absent in clinopyroxenites of this study.

A new suite or series must be proposed for rocks (xenoliths) related to processes of metasomatism. Thus, the nodules described here differ from typical representatives of the wehrlite-suite xenoliths recorded from all over the world; they are believed to be crystallization products of an alkali- and volatile-rich magma/fluid phase.

Two types of metasomatic processes, different in their chemistry, can be distinguished: 1) phlogopite-bearing rocks related to a mobile phase rich in K, Ti, Fe, and H_2O but also in P_2O_5 (apatite replacing phlogopite), and 2) alkaline pyroxenites: products of crystallization of a melt/fluid rich in Na, Ti, Fe, P, Cl, CO₂, and probably H_2O . The role of different elements in these two types was different: the most important for the first were K, H_2O , (Fe, Ti); for the second Na, CO₂, P, (Ti, Fe). The relations between these two types of metasomatic processes are unknown, but the fact that processes of mantle metasomatism must have operated can be regarded as established.

It remains open for discussion in what form these processes took place – was there a process of 'veining' of the upper mantle, a 'front of metasomatism', or infiltration of fluids in an other form.

Summary

The variety of types of xenoliths found in the Finkenberg basalts shows that the upper mantle beneath this volcanic region must be heterogeneous. It has been shown that the most abundant are xenoliths of spinel-lherzolites, thus the upper mantle is probably dominated by these rocks with subordinate websterites, harzburgites and dunites. The major mantle rock types are usually regarded as residua left after the processes of partial melting. Such rocks as the harzburgites and dunites of this study may belong to this group. Rocks of the lherzolite suite containing clinopyroxene as a major mineral phase, on the other hand, are thought to be fragments of mantle material, which have undergone a relatively small degree of melting. Thus, the undepleted mantle has here the composition of spinel-lherzolite and the generation of basaltic magmas must have taken place within the limits of the stability field of spinel (no garnet, no plagioclase).

Other types of xenoliths (wehrlite suite) can be considered crystallization products of a melt of basaltic or alkali-basaltic composition. Wehrlites and augite-bearing clinopyroxenites (both lacking phlogopite) can be regarded as cumulates which must have crystallized under relatively high-pressure conditions and be related to the host basalts or similar parental melts.

A special type of nodules is represented by rocks containing alkali- and volatile-rich minerals. They differ strongly in composition from any kind of mantle rock – fertile as well as depleted. The enrichment of mantle material in incompatible elements is generally thought to be related to the infiltration of volatile-rich fluids migrating upwards; their influx leads to the 'veining' of the upper mantle by hydrous minerals and to the creation

of a metasomatized mantle. These processes are widely developed and Boettcher et al. (1977) have stated that 50% of all nodules contain a noticeable amount of hydrous phases. The melting of the metasomatized mantle leads to the generation of melts rich in volatiles, alkalis and other incompatible elements.

Two groups of metasomatic processes have been noted in xenoliths of the Finkenberg; they differ from each other by their chemical particularities. The first one, represented by phlogopite-bearing rocks, is characterised by an important role of such elements as K, Fe, Ti, and H₂O. It is known from experimental work that under mantle conditions phlogopite can crystallize only from a H₂O-rich and CO₂-poor mobile phase. In the second group (apatite-bearing clinopyroxenites), the major active elements are Na, Fe, Ti, and also CO₂ and P₂O₅ (possibly also H₂O, but hydrous minerals are absent). The principal difference between these two groups is thus the different role of K and Na as well as of H₂O and CO₂. It is of particular interest that Dawson and Smith (1982) have stressed the 'contrasting type of metasomatism' in the upper mantle for different tectonic environments: potassium affinity for the subcratonic upper mantle and sodium affinity for regions of plate collisions and subduction zones. In the case of the Finkenberg these types (on a small scale) are found at one and the same place.

We have noticed what kinds of nodules occur together in one sample (Table 5). It can be seen that spinel-lherzolites occur together along with all types of xenoliths and also with xenocrysts and 'residual cores' of green clinopyroxene in the phenocrysts of basalts. Clinopyroxenites with green-clinopyroxene occur rarely with phlogopite-bearing rocks and the apatite-bearing clinopyroxenites have not been found together with any other nodules. These data have a restricted value but we have no other information about the spatial distribution of the different kinds of xenoliths.

All data mentioned above illustrate the existence of heterogeneities in the upper mantle of the Finkenberg region. In a relatively small area geochemically distinct processes in the mantle can be noticed: the partial melting with different degrees and different types of metasomatism. It is difficult to say whether these processes have taken place at different depth levels in the mantle, or whether they reflect another spatial distribution. It is also

sample (RGM)	spinel- lherzolites	wehrlite	phlogopite-bearing wehrlite, clinopy- roxenite	green pyroxene clinopyroxenite	xenocrysts of green clinopyroxene in basalts
318967	+			+	
318 984	+			+	
300 000	+		+		
319 008	+			+	
319 009		+		+	
319 025		+	+		
319 163	+				+
319 178	+				+
319 348	+				+
319 374	+				+
319 391	+				+
319 392	+		+		
319 392a	+	+	+		
319 396	+		+		
319 407	+		+		+

Table 5. Different types of xenoliths occurring together in the samples.

difficult to establish the succession of the two types of metasomatism. It seems reasonable to suppose the existence of small-scale heterogeneities, local processes of metasomatism and the presence of discrete magma reservoirs in the upper mantle. The processes of metasomatism and the partial melting of a metasomatised mantle (several metasomatic and melting events?) preceded the generation of the host basaltic magma; this latter process has probably had a larger regional development.

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Fig. 1. Lherzolite; porphyroclastic texture with relics of poikilitic texture; two types of olivine grains (OI) are present: small (included in pyroxene) and porphyroclasts; porphyroclasts of orthopyroxene (Opx), aggregates of small grains of clinopyroxene (Cpx); × 2.5, parallel polarizers.



Fig. 2. Lherzolite; relics of poikilitic texture: olivine (Ol) included by clinopyroxene (Cpx); × 40, crossed polarizers.





Fig. 1. Lherzolite; aggregate of desintegrated clinopyroxene (Cpx); × 40, parallel polarizers.



Fig. 2. Lherzolite showing spongy habit of clinopyroxene (Cpx); × 40, crossed polarizers.





Fig. 1. Lherzolite showing desintegration of clinopyroxene (Cpx); × 100, crossed polarizers.



Fig. 2. Lherzolite; localization of spinel grains (Sp) between pyroxene crystals; × 25, parallel polarizers.



- Fig. 1. Lherzolite with exsolution lamellae of spinel (two directions) in orthopyroxene; × 250, parallel
- polarizers.
 Fig. 2. Websterite; exsolution lamellae of spinel (two directions) in clinopyroxene grains oriented 1 [001]; cleavage; × 250, parallel polarizers.





Fig. 1. Websterite; porphyroclastic texture, porphyroclasts of orthopyroxene (Opx) and clinopyroxene (Cpx) in a matrix of clinopyroxene; × 10, crossed polarizers.



Fig. 2. Websterite; the form of spinel grains (Sp); \times 40, crossed polarizers.



Plate 5



Fig. 1. Phlogopite-bearing olivine-clinopyroxenite; poikilitic inclusions of olivine (Ol) and phlogopite I (Phl) in clinopyroxene (Cpx); large flakes of phlogopite II (Phl II) with equilibration relations to clinopyroxene; × 100, parallel polarizers.





Fig. 1. Phlogopite-bearing clinopyroxenite; newly-formed olivine (Ol) at the boundary of clinopyroxene (Cpx) and phlogopite (Phl) grains; spongy habit of clinopyroxene (sp-Cpx) can be observed; × 100, parallel polarizers.



Fig. 2. Wehrlite; apatite (Ap) and secondary phlogopite (Phl) replacing phlogopite II; \times 250, parallel polarizers.



Plate 8

- Fig. 1. Olivine-bearing clinopyroxenite; rim of a titanium-rich phlogopite on flakes of phlogopite II at the contact with basalt; × 250, parallel polarizers. Fig. 2. Apatite- and titanite-bearing clinopyroxenite; zonation and recrystallization of clinopyroxene;
- \times 40, crossed polarizers.

