Upper-mantle amphiboles: a review

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ABSTRACT. Textural and chemical data are reviewed for amphiboles occurring in Cr-diopside series and Al-augite series xenoliths; as megacrysts in a variety of igneous environments; and also as minor phases in tectonic slices of upper-mantle peridotite. New data are given for pargasite and associated phases in a lherzolite from Tanzania and in kelyphite replacing garnet in a S. African kimberlite xenolith. It is concluded that only pargasites and titaniferous pargasites occurring in Cr-diopside series blocks have formed under upper-mantle conditions. Although amphibole is present in the upper mantle, as suggested by Oxburgh (1964), its relative paucity suggests it is not a major alkali and water storage site; phlogopite is a more likely candidate, particularly in the deeper parts of the upper mantle.

THERE has been considerable interest in the origin and movement of volatile and incompatible elements within the upper mantle since Oxburgh (1964) suggested that amphibole might be present in the upper mantle to explain the disparity between the potassium content of mantle-derived basalts and primary parental peridotite. Because of its widespread occurrence in xenoliths transported by kimberlites, phlogopite has become regarded as the dominant volatile-bearing phase within the upper mantle (e.g. reviews by Delaney et al., 1980; Smith et al., 1979; Boettcher et al., 1979). None the less, over the years, reports have accumulated of amphibole in parageneses thought to be derived from the upper mantle. Amphiboles have now been found in peridotitic or pyroxenitic xenoliths, in megacryst suites of basic and ultra-basic volcanic rocks, and in upper-mantle lenses tectonically emplaced within crustal terrains. We review the variety of amphiboles found in these environments in order to complement the earlier reviews on phlogopite. The evidence for origin in the upper mantle rather than the crust is reviewed in the Discussion section.

Amphiboles in ultramafic blocks in basic or ultra-basic lava flows, in scoria cones or volcanic breccia vents. The xenoliths are commonly of two types.

One comprises rocks rich in forsteritic olivine (generally greater than 60% by volume) with subordinate enstatite or bronzite with an emerald green chromiferous diopside; small amounts of chrome spinel or pyrope garnet are the principal aluminous phases. A second type usually varies greatly in modal amounts of olivine, black titaniferous augite, and spinel, all of which contain relatively large amounts of TiO_2 and Al_2O_3 but lower Cr_2O_3 than equivalent phases in the first group; with the addition of plagioclase these rocks may grade into gabbros. Various names are used such as the Cr-diopside and Al-augite groups (Wilshire and Shervais, 1975) or groups 1 and 2 (Frey and Prinz, 1978) respectively. The terms *lherzolite series* and *wehrlite series* (Aoki and Kushiro, 1968; Kuno, 1969) can be implicitly confusing because lherzolite and wehrlite *modes* can occur in both series.

Occurrences of amphibole in ultramafic blocks from volcanic environments are listed in Table I. The list is not claimed to be comprehensive, but does illustrate the world-wide distribution and the variety of tectonic environments. The xenoliths were transported from depth by a variety of ultrabasic and basic rock types, which are silica-undersaturated except for the Kamchatka example; some (e.g., kimberlites and alnöite) show extreme alkali and volatile enrichment. The host rocks are generally basic or ultra-basic, except for the Tenerife phonolite. The host rocks occur in a variety of tectonic settings. The South African kimberlites and the examples from the Colorado Plateau and Arizona are on ancient, faulted cratons. The occurrences in Japan, Kamchatka and Alaska are at plate-collision margins; whereas the South Yemen and East African examples are in a crustal-tension environment. The only new sample in Table I is from the Engaruka area, N. Tanzania, given to the authors by Dr C. H. Emeleus; it was collected by D. Fairhead from an unidentified tuff

cone south of Lake Natron, in an area characterized by numerous small tuff cones, tuff rings and explosion craters (Dawson and Powell, 1969).

In the Cr-diopside series, pargasite and titanian pargasites are the commonest amphiboles, while potassium richterite occurs in strongly metasomatized harzburgites and lherzolites in some South African kimberlites. Some pargasites and titanian pargasites appear to be in textural equilibrium as grains of equal size to the others in the rocks or

TABLE I. *Occurrences of amphibole-bearing blocks*

Locality	Xenolith Type	Type of Amphibole	Host Rock	Amphibole in Megacryst Suite	References
Wesselton, South Africa	Harzburgite	Pargasite	Kimberlite	—	Boyd (1971)
Monastery, South Africa	Harzburgite	Pargasite	Kimberlite	—	Dawson and Smith (1975)
Hebron, South Africa	Gt. lherzolite, lherzolite	Pargasite	Kimberlite	—	Robey (1981)
Bultfontein, South Africa	Harzburgite, lherzolite	K-richterite	Kimberlite	—	Erlank and Rickard (1977); Dawson (1979)
Monastery, South Africa	Harzburgite	K-richterite	Kimberlite	—	Erlank and Rickard (1977)
De Beers, South Africa	Harzburgite	K-richterite	Kimberlite	—	Erlank and Rickard (1977)
Roberts Victor, South Africa	Eclogite	Hornblende	Kimberlite	—	Bishop et al. (1978)
Roberts Victor, South Africa	Eclogite	K-pargasite	Kimberlite	—	Boettcher et al. (1979)
Lovedale, South Africa	Eclogite	Edenitic hornblende	Kimberlite	—	Robey (1981)
Lovedale, South Africa	Kyanite eclogite	Pargasitic hornblende	Kimberlite	—	Robey (1981)
Stockdale, Kansas, USA	Eclogite	Titanian pargasitic hornblende	Kimberlite	—	Meyer and Brookins (1971)
Bultfontein, South Africa	MARID	K-richterite	Kimberlite	—	Aoki (1975); Dawson and Smith (1977)
Navajo Province, USA	Inclusion in garnet	Pargasite	Kimberlite	+	Hunter and Smith (1981)
Bellsbank, Jagersfontein, South Africa	Alkremite	Pargasite	Kimberlite	—	Nixon et al. (1978)
Malaita, Solomon Islands	Gt. orthopyroxenite	Titanian pargasite	Alnöite	—	Nixon and Boyd (1979)
Green Knobs } USA Buell Park	Lherzolites	Tremolite, richterite, edenite, pargasite	Minette	—	D. Smith (1979)
Dish Hill } Deadman Lake USA	Lherzolites	Pargasite, kaersutite	Basanite	—	Wilshire et al. (1980)*
Grand Canyon, USA	Lherzolite, pyroxenite	Kaersutite, richterite, pargasite	Basanite	+	Best (1974, 1975)
San Carlos, USA	Websterite, pyroxenite	Kaersutite	Basanite	+	Frey and Prinz (1978)
Nunivak, Alaska, USA	Lherzolite, gt. pyroxenite	Titanian pargasite, kaersutite, chromium pargasite	Basanite-alkali ol. basalt	—	Francis (1976a, b)
Hawaii, USA	Lherzolite, wehrlite, pyroxenite	'Hornblende', kaersutite	Alkali ol. basalt nepheninite	—	White (1966); Mason (1968); Kuno (1969)
Kamchatka, USSR	Lherzolite	Pargasitic hornblende	Andesite-basalt	—	Shcheka and Shcheka (1973)
Itinomegata, Japan	Gt. lherzolite, lherzolite, websterite	Pargasite, pargasitic hornblende	Alkali basalt	+	Aoki and Shiba (1973); Takahashi (1980)
Iki Island, Japan	Peridotite, pyroxenite	Ti-pargasite, kaersutite	'Basalt'	+	Aoki (1970)
Gloucester, NSW, Australia	Gt. clinopyroxenite	Magnesio-hornblende, ferroan pargasite	'Basalt'	+	Wilkinson (1974)
Barraba, NSW, Australia	Wehrlite	Ti-pargasite	Analcimite	+	Wilkinson (1975a)
Tumu-Ecumbene, NSW, Australia	Gt. pyroxenite	Pargasite	Basanite	+	Irving (1974a)
Victoria, Australia	Lherzolite, gt. wehrlite	Pargasitic hornblende	Basanite	—	Frey and Green (1974); Irving (1974a, b)
Kakanui, New Zealand	Gt. clinopyroxenite ('eclogite')	Kaersutite	Alkali ol. basalt	+	Dickey (1968); White and Chappell (1972); Merrill and Wyllie (1975)
Fife, Scotland	Ol. mica, clinopyroxene aggregates	Kaersutite, ferroan pargasite	Basanite/ monchiquite	+	Chapman (1976)
Fen, Norway	Lherzolite	Pargasite	Damkjernite		Griffin (1973)
Rosenberg, Germany	Peridotite, pyroxenite	Ti-pargasite, kaersutite	Basanite	+	Vinx and Jung (1977)
Eifel, Germany	Lherzolite	Pargasite	Basanite	+	Stosch and Seck (1980)
Szigliget, Hungary	Lherzolite	Ti-pargasite	Alkali basalt	—	Embey-Isztin (1976)
Massif Central, France	Lherzolite	Pargasite	Alkali basalt	—	Brown et al. (1980)
Tenerife, Canary Islands	Kaersutite pyroxenite	Kaersutite	Phonolite, alkali basalt	+	Borley et al. (1971)
Ataq, South Yemen	Lherzolite	Pargasitic hornblende	Alkali basalt	—	Varne (1970)
Lashaine, Tanzania	Olivine pyroxenite	Ti-pargasite, Ti-pargasitic hornblende	Carbonatite/ ankaramite	—	Dawson and Smith (1973); Pike et al. (1980)
Engaruka area, Tanzania	Lherzolite	Pargasite	Not known	?	This paper

* Summarizes much earlier literature from these localities.
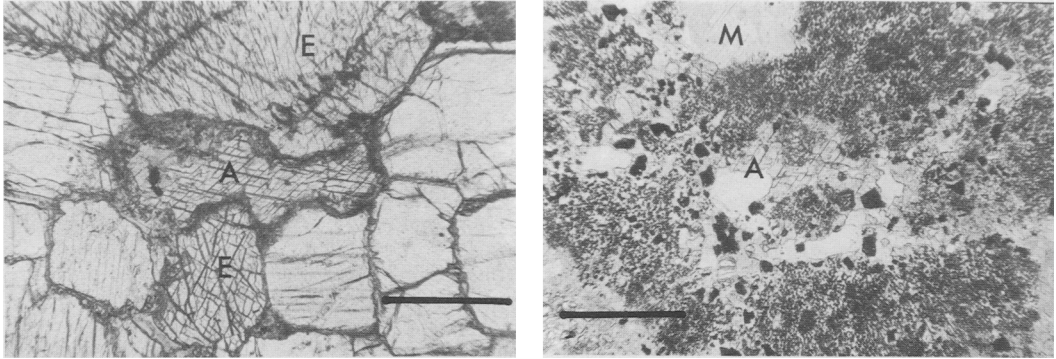
FIGS. 1 and 2. FIG. 1 (*left*). Pargasite (A) with enstatite (E) and forsterite in amphibole harzburgite xenoliths BD 1368, Monastery kimberlite, S. Africa. The pargasite contains small grains of chromite. FIG. 2 (*right*). Pargasite (A) in a 'kelyphite' aggregate replacing garnet in (former) garnet lherzolite BD 3707, Bultfontein kimberlite, S. Africa. Phlogopite (M) and spinel are visible; fine-grained diopside, phlogopite and pargasite form the matrix of the aggregate. Both figs. in plane-polarized light, scale-bar 0.5 mm.

as smaller interstitial grains (e.g. Varne, 1970). The pargasites in some harzburgites from South African kimberlites (Boyd, 1971; Dawson and Smith, 1975) (fig. 1) are also in equilibrium with surrounding phases. Pargasites which are intergrown with 'finger-print' chromite may result from metasomatism of an anhydrous silicate (pyroxene?) that grew simultaneously with the chromite. In other examples, there is abundant textural evidence that the amphibole is a later phase replacing earlier minerals. Robey (1981) reports that pargasite replaced both clinopyroxene and garnet in garnet lherzolite from the Hebron kimberlite, South Africa; amphibole replacing clinopyroxene has higher Na_2O and TiO_2 but lower Al_2O_3, Cr_2O_3 and CaO than amphibole replacing garnet.

Amphibole also occurs in 'kelyphite' replacing garnet in a lherzolite (BD 3707) from the Bultfontein Mine, where it forms a minor phase in clinopyroxene–phlogopite–aluminous spinel aggregates (fig. 2). Peridotite blocks in the Green Knobs kimberlite, New Mexico, contain replacement amphiboles whose compositional range from pargasite to tremolite correlates with the presence or absence of additional hydrous phases including antigorite and chlorite (Smith, 1979). Secondary hornblende occurs, with phlogopite, chlorite, calcite and ?devitrified glass, along grain-boundaries in some eclogites from the Roberts Victor kimberlites, S. Africa (fig. 3).

For the websterite and pyroxenite blocks, of the wehrlite series there is considerable debate whether
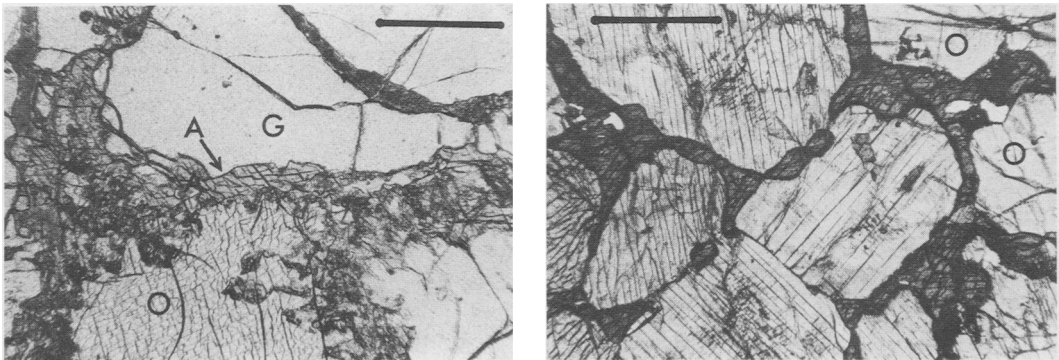

FIGS. 3 and 4. FIG. 3 (*left*). Secondary hornblende (A) at contact between garnet (G) and omphacite, in eclogite xenolith BD 1188, Roberts Victor kimberlite, S. Africa. FIG. 4 (*right*). Dark titanian pargasite in grain boundaries between diopside and olivine (O) in wehrlite block BD 804, Lashaine, N. Tanzania. Both figures in plane-polarized light, scale bar 0.5 mm.

the amphiboles (generally kaersutites) are igneous or metasomatic. In examples from Colorado, Tanzania, and Australia, the amphibole is interpreted as consolidated intercumulus liquid, with the olivine and clinopyroxene (fig. 4) being regarded as the cumulus phases (Best, 1970; Dawson and Smith, 1973; Irving, 1974a). The composition of the 'intercumulus' amphibole approximates to that of nephelinitic liquids and has been interpreted as the residual fluid in various deep-seated crystal fractionation models (e.g. Frey and Prinz, 1978). A cumulate origin is also proposed for kaersutite-apatite xenoliths from Kiama, Australia (Wass, 1979). Alternatively, the 'intercumulus' amphiboles in xenoliths from Dish Hill and Deadman Lake, California, have been attributed to infiltration metasomatism and replacement of original peridotite phases alongside kaersutite veins; amphibole composition in the wall-rock varies systematically from Ti-pargasite adjacent to veins, to pargasite at distance up to 8 cm from the veins (Wilshire et al., 1980). Whereas such an interpretation is incontrovertible for specimens adjacent to observed amphibole-rich veins, textural interpretation is more speculative for vein-free specimens.

The Californian specimens serve to introduce another group of small xenoliths in which amphiboles of various compositions comprise a high proportion of veins intruding into peridotite wall-rock. In the lherzolite block from Szigliget, Hungary (Embey-Isztin, 1976) amphibole is the major phase in cross-cutting veins; the only additional phase is a magnesian picotite. The veins are clearly intrusive though they have suffered recrystallization after consolidation. The amphibole-bearing samples from Kamchatka (Shcheka and Shcheka, 1973) comprise fragments of lherzolite that are commonly dissected by stringers of enstatite with chromite and, at some points, by 10–12 cm wide veins of coarse-grained enstatite with pockets of chromite replaced in the core or in strips by coarse grained diopside; chrome-bearing edenite was found in one of these wider veins which are interpreted as being of metasomatic origin. Amphibole-rich veins cutting peridotites in the Bultfontein kimberlite, S. Africa (Erlank and Rickard, 1977; Dawson, 1979) comprise varying proportions of K-richterite, phlogopite, diopside, and a K–Ba–titanate structurally related to crichtonite (Smyth et al., 1978; Jones et al., in prep.). Alongside the veins, the wall-rock phases have been very variably affected by the vein fluids, and enstatite is highly metasomatized, in contrast to olivine which is unaffected (fig. 5). In other peridotite blocks, not noticeably veined, from this locality, original diopside is partially replaced by richterite (fig. 6). Other compound xenoliths involving lherzolite in contact with hornblende-rich rocks were described from Kakanui, New Zealand (Dickey, 1968); here the commonest fragments are deformed lherzolite intruded or enveloped by undeformed hornblendite. In most specimens, the hornblendite–lherzolite contacts are sharp and show no reaction relationship, but in others the contact between the lherzolite and hornblendite is gradational. Subordinate grains of phlogopite and amphibole together with an apparently secondary chrome diopside were found in the lherzolite adjacent to the hornblendite veins, but no analyses were given of these phases.
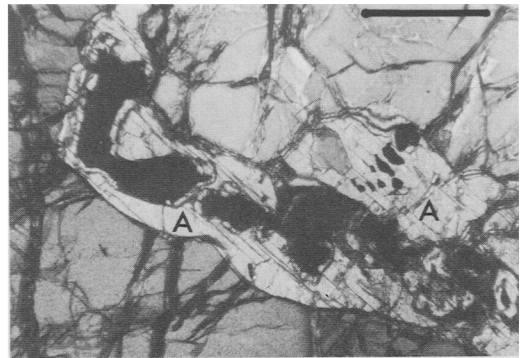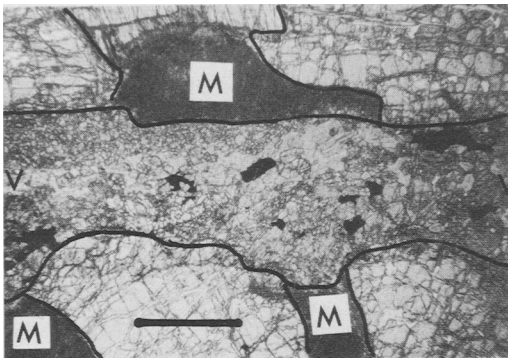
FIGS. 5 and 6. FIG. 5 (left). Vein (V), consisting of fine-grained richterite (higher relief), phlogopite and opaque K–Ba–titanate, cutting harzburgite block BD 3024, Bultfontein kimberlite, S. Africa. Wall-rock olivine (serpentinized) is unaffected but enstatite is replaced by fine-grained metasomatic aggregates (M) of phlogopite and richterite. Vein and grain contacts have been inked in for emphasis. Plane-polarized light, scale bar 1 mm. FIG. 6 (right). Richterite (A) replacing diopside, that has been turned into extinction for contrast; in lherzolite xenolith BD 3108 from the Bultfontein kimberlite, S. Africa. Crossed polars, scale bar 0.5 mm.

An additional feature observed in the Kakanui examples and not only in amphibole-rich veins but also in many amphibole-bearing peridotites, is the very common occurrence of Ti-bearing phlogopite associated with the amphibole. Most phlogopites appear to be in equilibrium with amphibole but the pargasites from the Monastery Mine (Dawson and Smith, 1975) and in some specimens from the Lashaine volcano (J. B. Dawson, unpubl. data) appear to have been replaced by later phlogopite.

In short, the amphiboles occurring in xenoliths of upper-mantle origin exhibit a variety of textures. Some have achieved textural equilibration with adjacent phases, whereas others occur interstitially and have been interpreted as consolidated inter-cumulus fluids: others undoubtedly result from replacement, in some cases with the replacing fluid being unambiguously connected with observed dyke-like intrusions which are rich in amphibole and, or, phlogopite. Analogies can be drawn between the textural features seen in this last grouping with those seen adjacent to mica–amphibole veins in some orogenic peridotite massifs such as Lherz, which will be discussed below.

Amphibole chemistry. Amphiboles in ultramafic blocks from the upper mantle vary considerably

in composition (Table II). Analyses of other phases occurring with the newly-analysed amphiboles from Tanzania and Bultfontein are given in Tables III and IV. The amphiboles range from pargasites

TABLE III. *Analyses of phases occurring with pargasitic hornblende in lherzolite DU18500, Tanzania (analysis 7, Table II)*

	Olivine	Cpx	Chromite
SiO_2	42.4	55.0	0.4
TiO_2	n.a.	0.14	0.8
Al_2O_3	n.a.	1.4	7.9
Cr_2O_3	0.03	2.0	60.8
FeO	8.1	2.3	17.9
MnO	0.16	0.09	0.20
MgO	51.5	17.2	12.9
NiO	0.27	0.04	0.10
CaO	0.11	20.7	0.19
Na_2O	n.a.	1.6	n.a.
K_2O	n.a.	0.01	n.a.
	102.57	100.48	101.19

Analyst: R. Hervig—analyses to one decimal place from solid-state detector, detection 0.1 wt. %, to two decimal places from spectrometers, accuracy 0.02 wt. %.

TABLE II. *Selected analyses of amphiboles in ultramafic blocks (xenoliths)*

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17
SiO_2	45.7	45.40	43.51	44.73	48.28	40.25	49.0	53.6	55.4	54.2	43.18	41.35	41.5	43.3	40.4	43.2	41.8
TiO_2	0.01	0.42	0.49	0.29	0.44	4.21	0.9	0.07	0.46	0.64	1.67	4.58	3.4	1.8	4.7	4.9	0.2
Al_2O_3	11.1	11.71	14.74	12.58	10.33	14.20	7.5	4.01	0.99	1.05	14.87	14.62	13.2	15.4	14.7	13.2	15.8
Cr_2O_3	2.10	1.53	1.07	2.43	1.79	0.0	2.1	1.16	0.48	0.11	1.18	0.4	0.25	1.13	0.1	n.d.	1.2
FeO	2.64	3.34	4.00	2.37	3.35	9.15	2.8	2.51	2.25	5.50	4.40	11.34	8.5	10.2	7.9	5.8
MnO	0.18	0.07	0.09	0.11	0.006	0.08	0.06	—	—	0.03	0.10	0.11	n.d.	n.d.	0.1	0.08	0.2
MgO	11.1	18.96	19.18	19.17	20.83	12.59	20.8	22.4	22.9	18.5	18.0	12.79	15.0	17.3	13.6	15.8	17.2
NiO	0.36	—	—	—	0.14	—	0.04	—	—	—	—	—	—	—	—	—	—
CaO	20.2	9.85	10.70	10.95	10.98	10.29	9.6	6.98	6.74	6.76	10.57	9.37	10.9	10.8	10.4	11.3	10.9
Na_2O	3.20	4.31	3.93	3.84	1.75	2.17	4.1	4.92	3.82	3.24	3.09	3.28	2.8	3.5	2.7	3.1	2.7
K_2O	1.34	1.20	0.19	0.43	0.19	1.99	1.0	2.35	4.63	5.25	1.63	1.37	1.3	0.02	1.5	0.84	1.6
	97.93	96.79	97.90	98.00	99.85	96.92	97.90	98.00	97.67	95.28	98.69	98.21	96.85	97.95	98.4	100.32	97.4

FeO is total Fe except analysis 5.

1. Pargasite, harzburgite, Monastery kimberlite, South Africa (Dawson and Smith, 1975).
2. Pargasitic hornblende, replacing garnet, garnet lherzolite 07183, Hebron kimberlite, South Africa (Robey, 1981).
3. Pargasite, lherzolite, Fen, Norway (Griffin, 1973).
4. Pargasitic hornblende, lherzolite, Ataq, South Yemen (Varne, 1970). Total includes 1.10 % Fe_2O_3.
5. Pargasitic hornblende, replacing diopside, enstatite, and chromite in pyroxenite vein cutting lherzolite block, Avacha volcano, Kamchatka (Shcheka and Shcheka, 1973). Total includes 0.06 % Fe_2O_3, 1.58 % H_2O^+, 0.12 % F.
6. Titanian pargasite, in amphibole-rich vein, cutting lherzolite block, Szigliget, Hungary (Embey-Isztin, 1976). Total includes 1.99 % H_2O^+.
7. Pargasite in lherzolite block from unidentified tuff-ring, north of Engaruka, Tanzania, Durham University collection number DU18500. New analysis by R. L. Hervig, Chicago.
8. K-richterite, peridotite in kimberlite, Monastery Mine, South Africa (Erlank and Rickard, 1977).
9. K-richterite, average of twenty-two samples, peridotite in Bultfontein kimberlite, South Africa (Erlank and Rickard, 1977).
10. K-richterite, in MARID nodule, Bultfontein kimberlite, South Africa (Dawson and Smith, 1977).
11. Ti-pargasite, replacing spinel in lherzolite 14001, Nunivak, Alaska (Francis, 1976a).
12. Kaersutite in gt. clinopyroxenite adjacent to lherzolite 14001, Nunivak (Francis, 1976b).
13 and 14. Pargasite in vein and interstitial titanian-pargasite, peridotite xenolith, Dish Hill, California (Wilshire et al., 1980).
15. Poikilitic kaersutite, lherzolite, Grand Canyon (Best, 1974).
16. Kaersutite, wehrlite, San Carlos (Prinz and Nehru, 1969).
17. Pargasite in 'kelyphite' replacing garnet, lherzolite BD3707, Bultfontein, South Africa. Average of 3 EDS analyses. Analyst: A. Jones, Chicago.

TABLE IV. *Analyses of phases occurring with pargasite in kelyphite in lherzolite BD3707, Bultfontein (analysis 17, Table II)*

	1	2	3	4	5
SiO_2	50.1	42.4	0.8	0.6	0.9
TiO_2	0.2	3.42	1.0	0.0	0.0
Al_2O_3	6.0	17.5	30.8	54.6	53.1
Cr_2O_3	0.6	1.47	30.6	6.6	10.9
FeO^T	4.0	4.42	19.7	17.2	14.5
MnO	0.3	0.02	0.7	0.0	0.0
MgO	15.5	18.0	14.7	19.6	18.9
NiO	0.0	0.02	—	—	—
CaO	20.0	0.00	—	—	—
Na_2O	0.6	0.41	—	—	—
K_2O	0.0	10.2	—	—	—
Sum	98.3	98.10	98.5	98.6	98.3

Analyst: A. Jones, Chicago microprobe.

1. Clinopyroxene—average of 3 EDS analyses.
2. Phlogopite, WDS analysis; sum includes 0.21 wt. % BaO, 0.03 wt. % Cl.
3. Brown spinel, average of 2 EDS analyses.
4, 5. Green spinel grains, 2 EDS analyses of each.

and chrome-bearing pargasites through titaniferous pargasite to kaersutite (the nomenclature of Leake, 1978, is used throughout this paper). In general, pargasite, pargasitic hornblende and titaniferous pargasite are most common in peridotites (the Cr-diopside series) whereas kaersutite is common in pyroxenite and websterite inclusions of the Al-augite series that have lower Mg/Fe ratios overall than the peridotites: there is an overall trend towards higher Fe, Ti, and Al in the amphiboles in the Al-augite series. The rather limited number of analyses of Fe_2O_3 (Table II) suggest quite variable Fe^{2+}/Fe^{3+}. In addition, in both peridotitic and pyroxenitic parageneses, the amphiboles are rich in Na though with varying Na/K ratios, and in the pargasites of replacement origin the Cr_2O_3 content of the amphibole varies according to the type of replaced mineral. Although Erlank and Rickard (1977) suggested that alkali metasomatism of lherzolite, with elimination of garnet, results in the formation of potassium richterite, the replacement amphiboles reported by Robey (1981) and here are pargasites: the one in BD3707 is very low in SiO_2 (41.8 wt. %) and its reasonably high Al_2O_3 and Cr_2O_3 contents are presumably inherited from the original garnet. Two further points of interest regarding the other phases in 3707 (Table IV) are that the kelyphite also contains (i) compositionally-different spinels with widely varying Al_2O_3, Cr_2O_3 and Mg/Fe and (ii) a highly-aluminous clinopyro-

xene. Similar spinels and clinopyroxenes were reported in anhydrous reaction coronas around garnets in lherzolite xenoliths from the Letseng kimberlite, Lesotho (Lock and Dawson, 1980). Furthermore Reid and Dawson (1972) suggested that 'kelyphitization' is a two-stage process, resulting from (i) an isochemical reaction between garnet and olivine to give Al-orthopyroxene + Al-clinopyroxene and spinel, attendant upon a change from garnet-peridotite to spinel-peridotite facies; followed by (ii) a metasomatic event, giving hydrous phases at the expense of the aluminous pyroxenes. If this two-stage process pertained in 3707, it appears that Al-orthopyroxene is more susceptible to metasomatism than clinopyroxene. In its relatively high TiO_2, Al_2O_3 and Fe/Mg the kelyphite phlogopite can be most closely matched with secondary-textured micas in lherzolite xenoliths and the high Cr_2O_3 is probably inherited from the garnet (Delaney et al., 1980).

Returning to the discussion of amphiboles, the richterites in MARID nodules (i.e. kimberlite-derived xenoliths consisting of mica, amphibole, rutile, ilmenite, and diopside) and in metasomatized and veined peridotites are K-rich, with K_2O greater than Na_2O. They are less titaniferous than kaersutite and less chromiferous than pargasites, and their very low Al_2O_3 matches that of the co-existing phlogopites. Although these potassic amphiboles are found in xenoliths that have been transported to the earth's surface by a K-rich medium (i.e. kimberlites), there is no genetic relationship between the xenoliths and the transporting igneous rock because of the quite different dates on the host rocks and xenoliths; however both may be manifestations of a K-rich upper mantle beneath southern Africa or a particular type of process for K concentration.

Values for volatile elements, trace elements and certain isotope ratios for xenolith amphiboles are shown in Table V, together with data for amphiboles in the orogenic peridotite environment. Rb values are variable which, together with varying K contents, give a wide range of K/Rb ratio: those in the Ataq pargasites are lower than in the pargasites from Nunivak which are also more variable, possibly reflecting their replacement origin. The Nunivak pargasites are relatively depleted in rubidium and in this respect resemble the kaersutites. Again, compared with the Nunivak pargasites and the kaersutites, the Ataq pargasite has both higher total strontium and higher $^{87}Sr/^{86}Sr$ ratios. On the basis of the limited data available, it appears that the amphiboles as a group are not particularly depleted in water and the $\delta^{18}O$ values are between 5 and 6‰ which is consistent with other upper-mantle phases (Dawson, 1981).

Values of δD vary considerably, being relatively high in the pargasites and low in richterites: the extremely low values for the MARID richterites may result from co-existence with earlier phlogopite which concentrated deuterium relative to residual water, leaving a lower value for the later-crystallizing amphibole (Kuroda *et al.*, 1977). Moreover Kuroda *et al.* (1977) suggested that the high F/(F + OH) in richterite relative to co-existing phlogopite points to a relative water deficiency at the time of amphibole formation. In addition to the data listed in Table IV, *REE* data were given by Varne and Graham (1971) for Ataq pargasite; the amphibole contains 97 ppm total *REE* and is highly enriched in the light *REE* with a La/Yb ratio of 47. Kaersutite from a Kakanui eclogite contains 55 ppm *REE* and, like the Ataq pargasite, is enriched in light *REE* with a La/Yb ratio of 10.7 (White and Chappell, 1972).

Megacrysts. Large crystals of amphibole up to 3 or 4 cm in length are found in scoria cones, volcanic vents, sills, and lava flows in many volcanic areas throughout the world. Some well-investigated localities (Table I) demonstrate that many megacrysts occur in association with xenoliths of the Al-augite series. The host rocks are generally silica-undersaturated basic or ultrabasic rock types such as nephelinite or basanite. Rounding of some megacrysts indicates corrosion by the host magma, whereas others exhibit good crystal faces and were either in equilibrium with the host rock or were transported too rapidly for corrosion;

this underlines the need for using megacryst as a purely descriptive word since some may be xenocrysts (see below) whereas others may be high-level phenocrysts with good crystal shapes. At many localities the amphiboles form part of a megacryst suite in which the other megacrysts are titaniferous phlogopite, Tschermakite-rich augite, apatite, albite and anorthoclase (e.g. Dickey, 1968; Irving, 1974c; Wilkinson, 1975a; Chapman, 1976), and at some localities there are aggregates consisting of varying proportions of megacryst phases. At Kakanui, New Zealand (Dickey, 1968), and Elie Ness, Fife (Chapman, 1976), titanian pyrope is an additional megacryst phase and at Kakanui forms aggregates with clinopyroxene to form 'eclogite'. Orthopyroxene is also reported in the megacryst suite at Grand Canyon (Best, 1970) and Anakies, Victoria (Irving, 1974c).

Selected analyses of megacrysts are shown in Table VI. High amounts of Ti and, conversely, extremely low amounts or absence of Cr_2O_3 seem to be characteristic of these amphiboles. Kaersutite (Leake, 1978) is the best name for most of them, and perhaps titaniferous pargasite for those with lower TiO_2. The crystals are chemically and optically homogeneous with no signs of zoning, though at any individual locality, the kaersutites may show considerable variation with respect to Mg and Fe, suggesting some fractionation (Irving 1974c; Vinx and Jung, 1977).

Their relatively large size has encouraged detailed investigations into amphibole chemistry.

TABLE V. *Volatile, trace element and isotope data for amphiboles in ultramafic blocks and orogenic peridotites (Finero and Lherz)*

		Rb ppm	Sr ppm	Ba ppm	K/Rb	$^{87}Sr/^{86}Sr$	F%	H_2O%	$\delta^{18}O$‰	δD‰	Ref.
Pargasite	Nunivak lherzolites	3.7– 4.5	600– 800	nd	1200–280	0.7027–0.70337					1
	Ataq lherzolites	8.4–10.5	1100–1400	nd	330–470	0.7034–0.70408					1
	Itinomegata gt. websterites							1.73–1.88		−27.9–28.5	3
Hornblende	Eifel pyroxenite							1.52		−31.8	3
Kaersutite	Kakanui eclogite	7.8	480	387	1301	0.7030					5
	Dish Hill vein	9.5	600	243	1337	0.7027	0.03	0.68	5.57	−54	2, 5
	Eacham wehrlite						0.00	1.21	5.13	−49	2
Richterite	Kimberley MARID rocks						0.51–0.70	1.72–1.74		−103.0–132.0	3, 4
	Kimberley lherzolite						0.38	2.34		−79.5	3, 4
Pargasitic hornblende	Finero lherzolite	7.66	308	62	2104	0.7037					5
Kaersutite	Lherz vein	5.8	511	16.3	2260	0.7036	0.03	1.30	5.26	−87	2, 5

nd = not determined.

References: 1. Menzies and Murthy, 1980.
2. Boettcher and O'Neil, 1980.
3. Kuroda *et al.*, 1977.
4. Kuroda *et al.*, 1975.
5. Basu, 1978.

TABLE VI. *Selected analyses of megacryst kaersutites and titaniferous pargasites*

	1	2	3	4	5	6	7	8	9	10	11
SiO_2	40.00	40.10	41.2	39.50	40.2	39.87	40.28	40.0	39.6	42.1	37.4–40.0
TiO_2	5.67	4.44	4.16	5.40	3.98	4.43	4.85	4.3	5.4	2.9	4.1–6.2
Al_2O_3	14.64	14.55	13.5	14.29	12.7	14.71	14.53	13.6	12.9	11.6	13.3–15.4
FeO	11.70	10.05	9.55	6.88	9.71	8.89	9.84	11.0	10.2	5.9	8.8–15.4
MnO	n.a.	0.14	—	0.13	0.13	0.12	n.a.	n.a.	0.0	0.11	0.08–0.18
MgO	11.52	11.0	13.8	7.96	11.3	13.25	13.78	13.4	13.8	16.9	9.5–13.9
CaO	10.91	11.06	10.9	9.02	9.86	10.75	11.42	10.7	12.3	10.2	9.6–12.5
Na_2O	2.59	2.99	2.24	2.64	2.79	2.31	2.47	3.1	2.6	2.8	2.1–3.5
K_2O	1.36	1.62	2.17	2.12	1.79	2.39	2.16	1.7	1.3	1.7	0.8–2.0
	98.39	100.02	97.52	100.34	92.46	96.72	99.33	97.8	98.1	99.70	

Total Fe expressed as FeO, except analyses 2, 4, and 10.

1. Black Rock, Nevada (Best, 1970).
2. San Carlos, Arizona (Mason, 1968). Total includes 3.24% Fe_2O_3, 0.73% H_2O^+, 0.10% H_2O^-.
3. Nephelinite sill, Nandewar, NSW, Australia (Wilkinson, 1975b).
4. Anakies, Victoria, Australia (Irving, 1974c). Total includes 12.02% Fe_2O_3, 0.16% P_2O_5, 0.22% H_2O^+.
5. Kakanui, New Zealand (Dickey, 1968).
6. Elie Ness, Scotland (Chapman, 1976).
7. Rosenberg, Germany (Vinx and Jung, 1977).
8. Sauclieres, France (Mason, 1968).
9. Phonolite lava, Las Canadas, Tenerife (Borley et al., 1971).
10. Nephelinitic breccia, Moroto, Uganda (Varne, 1968). Total includes 3.5% Fe_2O_3, 1.9% H_2O^+, 0.09% P_2O_5.
11. Range of specimens from eleven localities (Kesson and Price, 1972); total iron as FeO, actual values Fe_2O_3 0.64–12.4; FeO 1.1–12.48; P_2O_5 0.07–0.17; H_2O^- 0.01–0.18; H_2O^+ 0.1–2.09; Cl 0.03 and 0.09; F 0.14–0.42. Also data on twenty trace elements.

Boettcher and O'Neil (1980) reported a range of values of 0.38–47.45 for Fe_2O_3/FeO ratios in kaersutite megacrysts and an inverse correlation between this ratio and H_2O^+; amphiboles with large concentrations of Fe^{3+} are poor in H_2O and presumably rich in oxygen. The range of $\delta^{18}O$ values for the megacrysts (4.65–6.01 ‰) resembles that for the xenolith amphiboles (Table IV). Values of δD for kaersutites (-30 to -60) demonstrate enrichment in deuterium relative to deep-seated pargasites, richterites in kimberlite xenoliths, and vein kaersutite from Lherz (Table V), and phlogopite megacrysts from kimberlites (Sheppard and Dawson, 1975; Boettcher and O'Neil, 1980); none the less these δD values resemble those for kaersutites from Al-augite-series xenoliths and a vein from Dish Hill. Garcia et al. (1980) determined that H_2O is the predominant volatile in amphibole megacrysts from the southwest USA and that it varies inversely with both F and Ti. Kaersutites from the United States, Australia, New Zealand, and France have extremely variable K/Rb, but most have very consistent $^{87}Sr/^{86}Sr$ ratios in the range 0.70246–0.70290. Kaersutite megacrysts have lower $^{87}Sr/^{86}Sr$ ratios than host alkaline basalts, suggesting that they are not genetically related, though three kaersutite megacrysts from Dish Hill, California, have the same $^{87}Sr/^{86}Sr$ ratio as kaersutite in a vein cutting an ejected peridotite block (Basu,

1978). This was taken as indicating that the megacrysts are derived by fragmentation of pre-existing vein material (Wilshire et al., 1980).

Amphiboles in alpine-type peridotites. Amphiboles occur as minor constituents of ultramafic rocks in orogenic zones. Broadly, there are two different types of such ultramafic rocks; (a) those formed by accumulation of mafic phases in either concentrically zoned or layered igneous bodies; and (b) lens-shaped bodies of peridotite that have faulted contacts with surrounding rocks and which are interpreted as being inthrust, detached bodies of original upper-mantle peridotite that have been tectonically up-thrusted during plate collisions. Hornblende was reported as an intercumulus phase in peridotite from the Finero complex, northern Italy (Cawthorn, 1975) and kaersutite was regarded as the product of an intercumulus liquid in wehrlite ultramafic rocks from the Mikabu zone, central Japan (Inomata and Tazaki, 1974) and the northern Kanto mountains, central Japan (Tazaki and Inomata, 1974). Because this type of complex is crystallized at crustal levels, as witnessed by the presence of plagioclase in associated gabbros, it will not be considered further, and readers are referred to the review by Cawthorn (1975).

There are few petrological descriptions that combine detailed textural and chemical information for mantle-derived tectonic slices, and further

data are badly needed, especially as the spatial relationships are better developed than for exotic xenoliths. Pargasite from the Vondergreen Hill peridotite south-west Oregon (Henry and Medaris, 1980) is a stable phase with olivine, clinopyroxene, orthopyroxene, and spinel in a porphyroclastic-textured peridotite. In a different tectonic environment, pargasite occurs in mylonitized peridotites from the St. Paul's rocks on the mid-Atlantic ridge (Melson et al. 1967); the pargasite was an original phase in some peridotites prior to the recrystallization that accompanied the mylonitization. Titanian pargasite occurs in amphibole lherzolite which forms irregular patches in spinel-bearing lherzolite at Caussou in the northern Pyrenees, France (Conquéré, 1971a). Two alternative hypotheses for its formation are that (a) the amphibole lherzolite formed by localized recrystallization of the spinel lherzolite in the presence of a fluid phase containing water, titanium, and alkalis at approximately 1100 °C and 7–8 kb, or (b) the titanian pargasite precipitated from a liquid of nephelinitic composition caused by limited partial melting of the spinel lherzolite. Close to Caussou is the more famous locality of Lherz where upper-mantle garnet peridotite is cut by simple and composite dykes and veinlets of kaersutite-bearing pyroxenite, websterite and lherzite; the lherzites are referred to as type LZA consisting mainly of strongly-deformed kaersutite and type LZB which are like LZA but with the addition of ilmenite and titaniferous phlogopite (Conquéré, 1971b). Garnet may or may not be present in these pyroxenites and lherzites. Conquéré (1971b) drew analogies between the compositions of the kaersutites and titaniferous pargasites at Lherz with those found in ejected blocks at such localities as Kakanui and Hawaii, the main difference being in their Mg/Fe ratios: and he also inferred that these ejected blocks may have been derived from upper mantle veinlets such as those at Lherz. This suggestion was also proposed by Embey-Isztin (1976) for the Hungarian composite xenolith, and by Wilshire et al. (1980) for the Dish Hill and Deadman Lake composite rocks; the same analogy could be made for the Kakanui composite xenoliths described by Dickey (1968). Conquéré pointed out that the bulk composition of the Lherz dykes varies from olivine tholeiite in anhydrous assemblages to olivine basanite, nephelinite, and melilite nephelinite in the more alkali- and hydrous-rich assemblages (lherzites); this indicates a differentiation process which might be due either to variation in the partial pressure of water over liquid derived from partial melting of an almost anhydrous peridotite or to segregation of residual liquids increasingly undersaturated and enriched in alkalis and Ti during

emplacement of the dykes. Because the pyroxenites (±amphibole and, or, mica) studied to date do not have the geochemical characteristics of liquids, and yet do occur as intrusive bodies with sharp contacts against wall rocks, some mechanical process of crystal–liquid segregation must be invoked. Wilshire et al. (1980) stated that the structural relationships within the dykes in Californian xenoliths indicate filter pressing from deformation which was coeval with the emplacement of the dyke systems. Although the amphibole and mica of some of these pyroxenite dykes have a poikilitic habit, the overall fabric is not like that in the wehrlite blocks of the Al-augite series for which a cumulate origin appears reasonable. Wilshire et al. (1980) also proposed that the bulk composition of each dyke rock was modified mainly by wall-rock reactions during emplacement, filter pressing during consolidation and partial re-melting following consolidation. Another differentiation mechanism, plating of liquid crystals on dyke walls, was suggested by several authors and reviewed by Irving (1980). In this model, crystals segregating from liquids within conduits become plated on to the dyke walls from the flowing magma, giving a variety of precipitates along a channel network depending on the depth, temperature difference between magma and wall rock, channel width, magma flow rate and possible interaction between the liquids and the wall rocks. However, this model does not explain why liquidus crystals migrate to the walls of the conduits rather than being concentrated within the centre of the flowing magma as was shown both by direct field observation (e.g. Drever and Johnston, 1967; Simkin, 1967) and experiment (Bhattacharji, 1967).

Widespread amphibolitization within the crust of original upper-mantle rocks was reported in the Seiad ultramafic complex of northern California (Loomis and Gottschalk, 1981). Hornblendes of variable composition in amphibolite or hornblende layers within dunite were interpreted to result from the introduction of hydrothermal solutions along fractures of deformed peridotites. Amphibole-hornblende-pyroxenite and pyroxenite bands formed by diffusion of components from the solution into the dunite. The absence of cumulus structures, the continuity of very thin layers, and cross-cutting relationships, suggest a long history of formation of the layers during the deformation history which did, none the less, take place in the crust rather than in the mantle.

Discussion. Because there may be a considerable temptation to assign any amphibole in an ultramafic paragenesis to the upper mantle, especially when present in xenoliths of the Cr-diopside suite, it is perhaps necessary to emphasize various difficulties. First of all, the lack of agreement on the

relationships between dykes and wall rocks, even within well-exposed massifs transported to the Earth's surface, should induce a sense of caution in proposing explanations of the petrography and mineral chemistry of exotic, veined xenoliths. Second, many amphiboles occur interstitially, and the pressure and temperature deduced from coarse anhydrous silicates may refer to an earlier stage prior to infiltration of volatile materials. This could lead to serious over-estimation of the depth of amphibole crystallization if the host rock had undergone upward transport before amphibole crystallization. There is inconclusive evidence that amphibole tends to form euhedral crystals even in metasomatic events, and great caution is needed in textural interpretation. Third, although many syntheses of amphibole have been made under controlled P and T, the stability field depends on so many chemical variables that it is currently difficult to pin down the stability limits for a particular composition. It is certain that amphiboles can grow from a wide range of basic and ultrabasic melt compositions over PT conditions consistent with both the upper mantle and the crust (e.g. Merrill and Wyllie, 1975, fig. 7), and it seems likely that at least most amphiboles cannot survive deeper than 80–100 km irrespective of the steepness of the geotherm (e.g. Wyllie, 1979, fig. 2). In particular, pargasite and kaersutite have a wide stability range (Holloway, 1973; Merrill and Wyllie, 1975; Mysen *et al.*, 1978), which extends to low P and T as evidenced by the type localities (Pargas, Finland, crustal marble; Kreutz, 1908; Kaersut, Greenland, alkali gabbro; Grössner and Spielberger, 1929). Hence the following discussion should be regarded as exploratory rather than dogmatic, especially in view of the evidence for complex tectonic events at crust–mantle interfaces (e.g. Medaris, 1981).

The current view of localities such as Lherz is that the hydrous dykes, such as lherzite, crystallized at pressures near 8 kb in the crust. Perhaps it might be argued that some of the composite xenoliths from the Al-augite series have also formed by infiltration of volatiles into the crust rather than the mantle.

For the K-richterites from the peridotite xenoliths from S. Africa, the textures and isotopic data definitely demonstrate that the veining and associated metasomatism took place prior to incorporation of the xenoliths into the transporting kimberlite (Erlank and Rickard, 1977; Jones *et al.*, in press). The paragenesis of the host harzburgites (forsterite, enstatite and Cr-spinel) is consistent with crystallization in the upper mantle, and the only uncertainty is whether the K-richterite might have crystallized after transport to a much higher level.

Because Kushiro and Erlank (1970) found K-richterite to be stable in garnet-free peridotite up to 30 kb and 1000–1100 °C, it will be assumed that all K-richterites from the Bultfontein xenoliths grew in the upper mantle.

Since the pargasite of BD3707 forms part of a kelyphite aggregate, it seems certain that it developed by metasomatic replacement of the alteration assemblage of garnet after the garnet lherzolite had moved into the stability field of spinel lherzolite (Reid and Dawson, 1972). Because kelyphitization seems to be essentially uniform across peridotite xenoliths, it might be argued that it occurred prior to transport of the xenoliths by kimberlite, rather than during the erratic thermal regime of the transportation event. Nevertheless, the interstitial position of the pargasite in 3707 (fig. 2) might be the result of late metasomatism associated with transport in the kimberlite rather than simply the result of metasomatism associated with the kelyphitization event.

Good candidates for origin in the upper mantle are the pargasites occurring with forsterite-rich olivine, enstatite, chrome-diopside, and chromite in the ejected blocks from Ataq, the Engaruka area of northern Tanzania, the Monastery mine, and the Eifel area of W. Germany. Calculations of PT conditions of formation are precluded by a lack of the appropriate combinations of phases. Although a crude estimate gives 900–1000 °C for equilibration in the Ataq lherzolite, the corresponding depth of 80–100 km for a steady-state continental geotherm (Clark and Ringwood, 1964) may be too high in this area of dilation and crustal spreading. Aoki and Kushiro (1968) suggested that the pargasite from the Dreiser and Weiher peridotites crystallized at 20–30 km depth, while Nixon and Boyd (1979) found titanian pargasite associated with pyropic garnet (> 16–20 kb?) in a garnet pyroxenite from Malaita.

For kaersutite, a considerable range of depths has been proposed. Best (1975) suggested ~ 65 km for the amphibole and phlogopite in the xenoliths from the western Grand Canyon where the amphibole is interstitial to cumulus minerals. Kaersutites from San Carlos have been assigned to either the lower crust (Prinz and Nehru, 1969) or the upper mantle (Mason, 1968), and those from the Iki islands of Japan were assigned to 20–30 km depth (Aoki, 1970). Crystallization within the crust was proposed for the kaersutites from east Fife, Scotland (Chapman, 1976), and Frechen (1972) suggested that the amphiboles from Dreiser Weiher crystallized from alkali-basaltic melts in a subvolcanic magma chamber. Best (1974) concluded that amphiboles from the mantle generally contain less Ti and more Al than those from the upper

crust, and Vinx and Jung (1977) therefrom proposed that the amphibole megacrysts from Rosenberg, Germany came from the mantle.

A 'chicken and egg' dilemma exists for kaersutite because of the contrasting propositions that melting of upper-mantle kaersutite can yield alkali olivine basalts and nephelinites (e.g. Varne, 1968; Forbes and Starmer, 1974), and conversely that kaersutite can crystallize from these types of melt (e.g. Conquéré, 1971b; Irving, 1980; Wilshire et al., 1980). The second proposition seems to be supported nicely by the evidence from many veined xenoliths, but the first proposition is probably too simple. The role of mica needs to be explored, and contributions from the anhydrous silicates can be expected. Currently it seems that only a few amphiboles can be assigned unambiguously to the upper mantle, and it will be necessary to develop reliable chemical and textural criteria before the ambiguous ones can be placed. Perhaps further experimentation will reveal distinctive chemical thermobarometers.

Finally, we draw attention to the contrasting types of veining and metasomatism in composite xenoliths from the sub-cratonic upper mantle in southern Africa, and in the ultramafic rocks from regions of plate collision (e.g. Pyrenees) and subduction of oceanic crust (e.g. western USA). In southern Africa, the liquids and their metasomatic effects are dominated by potassium, whereas in the other environments the dominant alkali is sodium. This contrast should provide useful information on the way in which the chemical heterogeneity of the upper mantle is related to the major tectonic forces involving the crust.

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