

*The significance of the gabbroic xenoliths from
Gough Island, South Atlantic*

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Summary. The petrography and mineralogy of a series of gabbroic xenoliths, consisting essentially of varying proportions of olivine, clinopyroxene, orthopyroxene, and plagioclase, are described from Gough Island. These are compared with xenoliths from other oceanic localities. Four new analyses of xenoliths and six of pyroxenes are presented, together with some partial analyses of plagioclases. It is concluded that the xenoliths are derived from the mantle above the magma source area. The composition of the mantle in this region is thought to be that of an olivine-tholeiite.

THE specimens described in this paper were collected by the author during the southern summer of 1955–56 while undertaking a general geological study of Gough Island. Although abundant xenoliths had been collected from both Tristan da Cunha and Ascension none had been recorded from Gough prior to this visit. The petrology of the volcanic rocks in which they occur has been described elsewhere (Le Maitre, 1962).

The xenoliths occur abundantly in many of the basalt and trachybasalt flows and dykes and also in tuff horizons, but do not appear to occur associated with any of the trachytic late differentiates. In shape they vary from rounded to angular and can be up to 20 cm across.

The term gabbroic is here used in its loose sense to imply a coarse grained rock containing, as its only feldspar, plagioclase more calcic than An_{50} . The major ferromagnesian minerals are olivine, clinopyroxene, and orthopyroxene. Although there is a considerable variation in the relative abundance of the constituent minerals from xenolith to xenolith little attempt has been made to use any of the more specific names of the clan e.g. norite or troctolite, as this would distract from some of the essential similarities between them. An exception, however, has been made in the case of the relatively rare peridotitic xenoliths in which plagioclase is virtually absent and the mafic minerals are more magnesian than in the gabbroic types.

Petrography

The *gabbroic xenoliths* have a granular to gabbroic texture in hand specimen and consist of black pyroxenes, up to 10 mm across, yellow-green to red-brown olivines and white to pink interstitial plagioclase. They are frequently friable.

In thin section they consist of augite, hypersthene, olivine, calcic plagioclase, and iron-ores, in widely varying proportions. The *augite* is a pale to medium grey-green often with fine lamellae and 'blebs' of exsolved hypersthene and occasionally rods and flakes of magnetite and dark brown ilmenite. Multiple twinning and inclusions of olivine and hypersthene also occur. When in contact with the lava, it invariably shows signs of reaction e.g. a granulated edge. The *hypersthene* (Fs_{25} – Fs_{30}), which is pleochroic from pale green to pale straw, frequently contains extremely fine exsolution lamellae of monoclinic pyroxene (parallel to the optic axial plane), but unlike the augite, it rarely has any other inclusions. Coarse intergrowths of hypersthene and augite are rare. Although usually stable in the xenoliths the hypersthene is always separated from the lava by a reaction rim of granular olivine which is frequently iddingsitized. However, this reaction sometimes takes place in the body of the xenolith where any adjacent plagioclase is usually converted to alkali feldspar. Large irregular grains of olivine are often associated with this reaction. The *olivine* (Fa_{23} – Fa_{28}) is anhedral and rarely fresh, the usual alteration being to 'iddingsite' (see Gay and Le Maitre, 1961). In the case of highly oxidized xenoliths, however, the olivine often exsolves hematite and a little magnetite, as hairlike rods and thin sheets. The *plagioclase* (An_{65} – An_{55}) usually occurs interstitially and is only slightly zoned. Large irregular grains of ilmenite and magnetite are common, and in some xenoliths, a brown basaltic amphibole is developing from the augite. Some of the more reactive xenoliths contain, often as a surface coating, a foxy-red biotite that is similar to that found in the trachybasalt lavas.

Occasionally the xenoliths have been completely recrystallized to a hornfels, the only difference in mineralogy being the scarcity of hypersthene.

The *peridotitic xenoliths* grade into the preceding type, as the amount of plagioclase gradually increases. They are rarely larger than a few centimetres across and are usually granular in appearance, consisting of small black pyroxenes and yellow-green to red-brown olivines.

In thin section the texture is gabbroic, but the *augite* is much paler in

colour than in the gabbroic xenoliths. In some xenoliths, however, it is distinctly yellow due to numerous minute orientated rods and 'blebs' of an exsolved yellow-green spinel. Exsolution lamellae of hypersthene are also present. The *hypersthene* ($\text{Fs}_{10}\text{-Fs}_{15}$) is also very pale with extremely weak pleochroism. Like the *augite*, this frequently has well-oriented inclusions of a yellow-green spinel. Fine exsolution lamellae of *augite* parallel to the optic axial plane also occur. The *olivine* (Fa_{10}) invariably shows marked signs of strain, roughly parallel to (100). It is usually fresh, except when it occurs in contact with the lava, where it alters to 'iddingsite'. Small crystals of brown spinel occur in minor amounts, but opaque iron ores are absent.

In contact with the lava these minerals exhibit some interesting reaction relations; hypersthene breaks down to iddingsitized olivine, while *augite* shows a slight change in composition and exsolves magnetite as opposed to the yellow-green spinel.

Less magnesian peridotitic xenoliths also occur and apart from the absence of plagioclase, they exhibit the same features as the gabbroic xenoliths.

Mineralogy

Olivines. In most cases their composition in the xenoliths does not differ by more than a few per cent. from those in the enclosing lavas (see table I). Where there is a marked difference, however, e.g. X10BB, X17AB, the xenolithic olivines are separated from the lava by rims of granular, often iddingsitized, olivine. In one extreme case, X18AF, the olivine (Fa_{10}) inside the xenolith is breaking down to granular hypersthene (Fs_{17}), which is in turn breaking down to granular iddingsitized olivine where it comes into contact with the lava.

Edwards (1938, p. 88) also records olivines breaking down to a granular rim of pyroxene (unspecified) in some of the basic xenoliths from Kerguelen.

Pyroxenes. Unlike the lavas, the xenoliths are often characterized by the presence of two pyroxenes, an orthorhombic hypersthene and a monoclinic *augite*. They are extremely variable in abundance and have only a small composition range (see table I).

A diagnostic feature of the xenolithic pyroxenes is the presence of exsolution of one pyroxene in another. These and other exsolution phenomena were investigated by single crystal X-ray methods (see Bown and Gay, 1959), which revealed five distinct types:

Augite exsolving hypersthene: This is in the usual orientation with the

TABLE I. Compositions and optical properties of some xenolithic minerals; and xenolith localities. Idd.: olivine altered to iddingsite. Refractive indices ± 0.002 , $2V \pm 2^\circ$. Compositions of olivine, hypersthene, and plagioclase inferred from values of β refractive indices

	Oliv. Mol. % Fa	Hyp. Mol. % Fs	Plag. Mol. % An	Augite β $2V$	
From trachybasalt boulder on beach, S. Buttress Bay.					
X10 rock*	28	none	48	1.705	48°
X10BB	17	22	n.d.	1.698	n.d.
From red tuff boulder on beach, S. Buttress Bay					
X11 rock	n.d.	n.d.	n.d.	n.d.	n.d.
X11A	23	rare	59	1.701	50°
X11B	23	26	61	1.700	50°
X11C	none	28	50	rare	
X11D	25	28	55	1.700	52°
X11E	rare	none	59	rare	
From trachybasalt boulder on beach, The Glen					
X15 rock	26	none	57	1.699	n.d.
X15C1	23	none	52	1.699	49°
X15C2	28	29	52	1.700	48°
X15CB	rare	27	n.d.	1.701	n.d.
X15F	27	none	54	1.700	49°
X15F1	24	none	61	1.700	48°
X15G	25	none	59	1.700	51°
From trachybasalt boulder on beach just south of Little Glen					
X17 rock	22	none	n.d.	n.d.	n.d.
X17AB	10	13	none	n.d.	n.d.
From basalt flow on beach just south of Little Glen					
X18 rock	24	none	54	1.701	n.d.
X18AB	25	27	59	1.701	n.d.
X18AC	25	27	none	1.702	n.d.
X18AF	10	10	none	1.682	52°
X18C	26	28	57	1.700	50°
X18E	24	none	54	1.702	51°
X18F	23	26	61	1.701	51°
X18G	Idd.	none	61	1.696	53°
X18H	20	25	none	rare	
From trachybasalt dyke in cliff 100 ft west of Dell Rocks					
G30 rock	24	none	54	1.703	n.d.
G30F	20	25	55	1.700	n.d.
G30P	22	26	none	n.d.	n.d.
G30S	rare	25	55	1.700	52°
G30T	25	27	55	1.699	50°
G30W	Idd.	27	54	1.701	n.d.

* This rock has been analysed (Le Maitre, 1962, table 10).

b and *c* crystallographic axes in common; this was confirmed optically in a coarse augite-hypersthene intergrowth in the xenolith X15A, in which the exsolution plane was (100) of the augite.

Augite exsolving magnetite and spinel: both the opaque magnetite and the yellow-green spinel have (111) and $(\bar{1}10)$ parallel to (100) and (010) of the augite respectively; the first of two possible orientations described by Bown and Gay. Optically these two minerals exsolve as rods parallel to the *c*-axis of the augite. The yellow-green spinel has a cell dimension $a\ 8.10 \pm 0.05\ \text{\AA}$ and would appear to be an aluminous variety (Vermaas and Schmidt, 1959, p. 224). The magnetite has $a\ 8.40 \pm 0.05\ \text{\AA}$.

Augite exsolving ilmenite: this is in the twinned orientation described by Bown and Gay, i.e. (0001) of the ilmenite parallel to (100) of the augite and either $(2\bar{1}\bar{1}0)$ or $(\bar{2}110)$ of the ilmenite parallel to the *c*-axis of the augite.

Hypersthene exsolving augite: there are two orientations with the *b*- and *c*-axes in common; in one the negative *a*-axis of the augite is inclined at approximately 16° to the positive *a*-axis of the hypersthene, and in the other the positive *a*-axis of the augite is inclined to the positive *a*-axis of the hypersthene. This is identical to the intergrowth described by Bown and Gay. Optically the lamellae are parallel to the (100) optic axial plane of the hypersthene; they are rarely wider than 0.002 mm and always disappear before reaching the edge of the crystal. It was thought that these might be misorientation lamellae similar to those described by Henry (1942) in plutonic and regionally metamorphosed hypersthene, as some of the single crystal X-ray photographs showed a double hypersthene pattern; however, as the patterns also showed reflections belonging to exsolved augite, which could have accounted for most of the lamellae, it was felt that this was unlikely and that the double pattern was due to multiple crystals.

Hypersthene exsolving spinel: there are two orientations with (111) of the spinel parallel to (100) of the hypersthene and either [100] or $[\bar{1}\bar{1}0]$ of the spinel parallel to the positive *b*-axis of the hypersthene. This yellow-green spinel also has a cell dimension $a\ 8.10 \pm 0.05\ \text{\AA}$.

The exsolution of iron-ores and spinels in the pyroxenes is undoubtedly due to reheating: hypersthene can be made to exsolve magnetite on heating in air for a few hours in the laboratory (Dr. P. Gay, personal communication).

Optical properties of some of the monoclinic pyroxenes are listed in table I. With the exception of the diopsidic augite (from the peridotitic

TABLE II. Chemical analyses and structural formulae of pyroxenes

	Hypersthene		Augite			
	G30S	G30T	G30S	G30T	X15C1	X18E
SiO ₂	51.88	52.8	49.25	50.2	49.9	49.4
TiO ₂	0.70	0.7	1.52	1.4	1.4	1.6
Al ₂ O ₃	4.24	2.2	5.26	4.2	4.5	4.7
Fe ₂ O ₃	1.13	1.7	2.27	2.3	2.1	2.5
FeO	12.69	12.6	6.40	5.9	7.0	5.8
MnO	n.d.	0.24	n.d.	0.12	0.14	0.13
MgO	25.39	26.1	15.08	15.6	15.0	15.0
CaO	3.37	3.4	19.14	19.6	19.0	19.9
Na ₂ O	0.25	0.2	0.78	0.8	0.5	0.8
K ₂ O	0.08	0.1	0.05	tr.	0.1	tr.
P ₂ O ₅	n.d.	0.03	n.d.	0.07	0.06	0.06
H ₂ O	0.04	0.10	0.04	0.04	0.20	0.10
Total	99.77	100.17	99.79	100.23	99.90	99.99
Structural formulae on basis of six oxygens						
Si	1.876	1.908	1.827	1.851	1.853	1.831
AlIV	0.124	0.092	0.173	0.149	0.149	0.169
AlVI	0.057	0.002	0.057	0.034	0.050	0.036
Ti	0.019	0.019	0.042	0.039	0.039	0.045
Fe ³⁺	0.031	0.046	0.063	0.064	0.059	0.070
Fe ²⁺	0.384	0.381	0.199	0.182	0.217	0.180
Mn	—	0.007	—	0.004	0.004	0.004
Mg	1.368	1.405	0.834	0.857	0.830	0.829
Ca	0.131	0.132	0.761	0.774	0.756	0.790
Na	0.018	0.014	0.056	0.057	0.036	0.057
K	0.004	0.005	0.002	—	0.005	—
Total	4.012	4.011	4.014	4.011	3.996	4.011
Ca	6.8	6.7	41.0	41.2	40.5	42.2
Mg	71.5	71.3	45.9	45.5	44.5	44.3
ΣFe	21.7	22.0	14.1	13.3	15.0	13.5

Analysts: G30S augite and hypersthene by R. W. Le Maitre;
remainder A. A. Moss.

xenolith X18AF, the monoclinic pyroxenes show very little variation in β (1.696 to 1.702) or $2V$ (48° to 52°) indicating a small composition range. In the highly oxidized tuff, X11, the xenolithic augites exhibit an increase of nearly 10° in $2V$ at the edges of the crystals, probably due to oxidation of iron (Segnit, 1953, p. 222). With the exception of X18AF and X17AB, there is also very little variation in the composition of the hypersthene (Fs₂₂ to Fs₂₉).

Four gabbroic xenoliths were selected for detailed chemical work; two contain clino- and orthopyroxene (G30T and G30S), while the other two contain only clinopyroxene (X15C1 and X18E). The analyses of

these six pyroxenes are given in table II. As can be seen there are no significant differences between the four clinopyroxenes, thus suggesting that the distribution of clino- and orthopyroxene in the xenoliths is governed by a random process rather than by some such mechanism as the unmixing, in situ, of a homogeneous sub-calcic pyroxene. This again suggests that the xenoliths are small fragments of a modally heterogeneous body composed of mineral phases of nearly constant composition.

These pyroxenes are plotted in fig. 1 together with other published pyroxene data for comparative purposes. From a study of this diagram alone there is little to suggest that the pyroxenes do not belong to a typical tholeiitic or nonalkaline assemblage: the clinopyroxenes are slightly poorer in lime than typical alkaline types and recalculated in terms of $(Ca + Na + K):Mg:(Fe^{3+} + Fe^{2+} + Mn)$ they fall in the field of nonalkaline pyroxenes as defined by Le Bas (1962, p. 268, fig. 1); the tie-lines between the two coexisting pyroxene pairs are in alignment with the tholeiitic tie-lines; and the composition of the olivines coexisting with the pyroxenes seems to be in accord with those from tholeiitic rocks, although the plagioclase is more sodic. Added to this is the fact that the presence of orthopyroxene as discrete crystals with, or as exsolution lamellae in, a clinopyroxene is usually regarded as sufficient evidence for designating an igneous mineral assemblage tholeiitic.

However, if the clinopyroxenes are classified by criteria involving Al, Ti, and Si (Le Bas, 1962; Kushiro, 1960) they are much more akin to alkali types. Likewise their norms show them, on the whole, to be Ne normative, a criterion that Yoder and Tilley (1962, p. 365) consider to be diagnostic of pyroxenes from critically undersaturated alkali magmas.¹ The clinopyroxenes, therefore, appear to fall somewhere between typically alkali and typically tholeiitic types.

Feldspars. The plagioclase, which is nearly always interstitial, is frequently multiply twinned. The most common types of twinning are albite and pericline, often in the same crystal. However, the areas of pericline and albite twinning are usually discrete, the twin lamellae wedging out at the boundaries of each area. Zoning, which is only slight, appears to be more common in the untwinned plagioclases. They are frequently crowded with minute 'dust' inclusions, iron ore, and small needles of apatite, especially in the recrystallized, hornfelsic xenoliths.

¹ It is only fair to point out that the most saturated clinopyroxene (X15C1 with Hy = 6.66 %) comes from the most undersaturated xenolith.

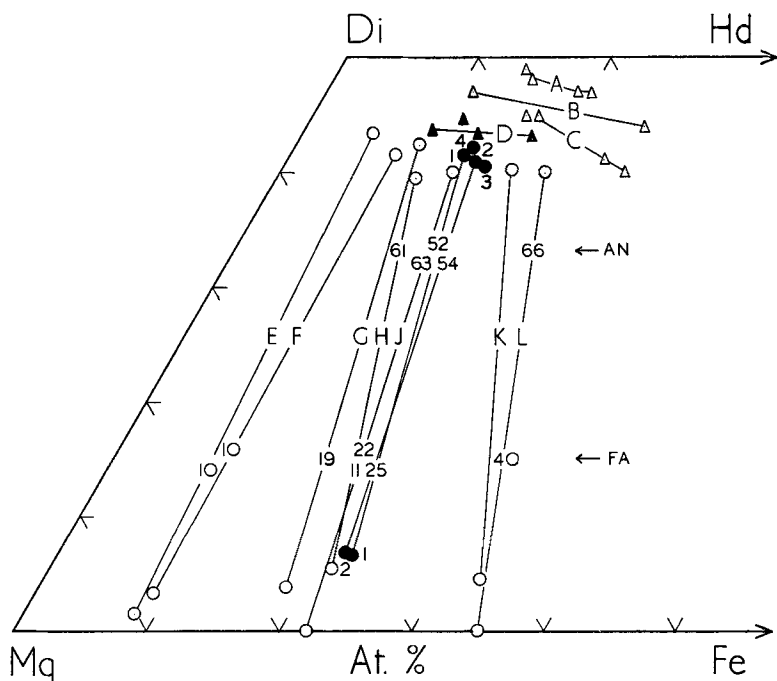


FIG. 1. Relationship between xenolithic pyroxenes (solid circles) and alkali pyroxenes in the lavas (solid triangles) from Gough Island, with other pyroxenes for comparison. 1 = G30T, 2 = G30S, 3 = X15C1, and 4 = X18E. Alkali pyroxenes (open triangles): *A* = Black Jack Sill, *B* = Garbh Eilean Sill, *C* = Morotu, Sakhalin, and *D* = Gough Island (for references see Le Maitre, 1962, p. 1318). Tholeiitic pyroxene pairs (open circles): *E* = peridotite nodule, Salt Lake Crater, Oahu, Hawaii; *F* = peridotite nodule, Kauai, Hawaii (Ross, Foster, and Myers, 1954, Amer. Min., p. 693, Anal. 8 and 9 respectively, tables 5 and 6); *G* = Gabbro-picrite, 4526, Skaergaard (Brown, 1957, Min. Mag., p. 511, Anal. 1 and 1a, table I); *H* = Uwekahuna gabbro, Hawaii (Muir, Tilley, and Soon, 1957, Amer. Journ. Sci., p. 241, Anal. 1a and 1b, table 4); *J* and *L* = Gabbros, Okonjeje (Simpson, 1955, Trans. and Proc. Geol. Soc. S. Africa, p. 125. Optically determined from specimens 7 and 1, figs. 2 and 3); *K* = Pitchstone, Sgurr of Eigg (Carmichael, 1960, Journ. Petrology, p. 309, Anal. 1 and 1a, table 2). An = Anorthite content of coexisting plagioclase, Fa = Fayalite content of coexisting olivine, where available.

The plagioclase in the xenoliths shows very little variation in composition compared with the total range in the lavas, ranging only from An_{61} to An_{50} (table I). This, however, is comparable with the range in composition shown by the cores of the phenocrysts in the basic lavas (Le Maitre, 1962, p. 1320).

The partial analyses of plagioclases from five of the xenoliths—four

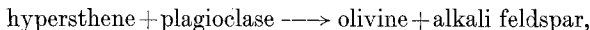
of them analysed—are given in table III together with a partial analysis of the plagioclase phenocrysts from a trachybasalt lava, G95, from Gough Island. As can be seen there is no significant difference between the phenocrystic and xenolithic plagioclases in terms of their major components Or–Ab–An. Both plagioclases X15F1 and G95 proved to be of intermediate to high structural state as determined by X-ray diffractometer methods outlined by Smith and Gay (1958).

TABLE III. Partial analyses and compositions of some plagioclases

	Xenolith plagioclase					Phenocrysts
	G30S	G30T	X15C1	X18E	X15F1	G95
Na ₂ O	4.98	5.02	5.24	5.30	4.84	4.70
K ₂ O	0.66	0.88	0.74	0.73	0.72	0.74
Or	3.8	5.1	4.3	4.2	4.1	4.2
Ab	43.6	43.9	45.8	46.3	42.4	40.4
An	52.6	51.0	49.9	49.5	53.5	55.4

Analysts: X15F1 and G95 by R. W. Le Maitre; remainder I. S. E. Carmichael.

In some of the more reactive xenoliths, e.g. X15C2, where hypersthene is breaking down to granular olivine in the centre of the xenolith, the adjacent plagioclase is often converted to an alkali feldspar with β 1.531 ± 0.003 . This reaction,



must involve exchange of material between the xenolith and the lava. Lime is undoubtedly removed and potash is probably added to the xenolith, although as the composition of the alkali feldspar is not known, the latter may be small. Silica may also be added or subtracted from the xenolith, depending upon the relative quantities of the reactants and products.

Iron–titanium oxides. The two main ores in the xenoliths are magnetite and ilmenite, although some of the highly oxidized xenoliths contain hematite and goethite. The *magnetite* is not abundant and occurs as anhedral crystals up to 1 mm across; occasionally it contains exsolution lamellae of ilmenite but more often it occurs as homogeneous grains, closely associated with the ilmenite, suggesting complete unmixing. The *ilmenite* occurs as abundant large anhedral crystals up to 2 mm across; coarse exsolution lamellae of magnetite are not uncommon and some of the crystals are also mantled by magnetite. In some of the highly oxidized xenoliths e.g. X11D, goethite occurs as individual

TABLE IV. Chemical analyses and C.I.P.W. norms of gabbroic xenoliths

	G30T	G30S	X15C1	X18E
SiO ₂	47.89	52.42	48.26	46.24
TiO ₂	1.19	0.68	1.27	3.42
Al ₂ O ₃	7.64	18.57	14.15	11.95
Fe ₂ O ₃	5.20	1.45	3.02	4.11
FeO	8.27	3.63	6.76	6.65
MnO	0.08	0.13	0.12	0.11
MgO	19.40	8.09	12.16	12.01
CaO	8.27	11.33	10.98	11.69
Na ₂ O	1.42	3.30	2.75	2.22
K ₂ O	0.36	0.45	0.39	0.39
P ₂ O ₅	0.11	0.05	0.07	0.13
H ₂ O ⁺	0.39	0.06	0.33	0.69
H ₂ O ⁻	0.19	0.07	0.10	0.46
Total	100.41	100.23	100.36	100.07
C.I.P.W. norms				
Or	2.13	2.66	2.31	2.31
Ab	12.01	27.92	22.54	18.78
An	13.41	34.53	25.12	21.49
Ne	—	—	0.40	—
Di	21.32	16.98	23.05	28.08
Ol	15.82	5.43	19.57	11.12
Hy	25.08	9.08	—	4.39
Il	2.26	1.29	2.41	6.50
Mt	7.54	2.10	4.38	5.96
Ap	0.26	0.12	0.17	0.31
% Fe in Femic				
	12.5	14.8	16.8	7.9
Or	7.5	4.0	4.4	5.3
Ab+Ne	45.1	44.3	47.4	45.6
An	47.4	51.7	48.2	49.1

Analyst: I. S. E. Carmichael.

grains and in fine veins. Hematite also occurs, either closely intergrown with ilmenite or in small veins.

Discussion

The analyses and norms of the four gabbroic xenoliths are given in table IV. Their heterogeneous nature is borne out by their bulk chemistry, which shows considerable variation in the major oxides, in particular MgO and Al₂O₃. These oxide variations appear in the norm as variations in total salic and femic constituents. It is interesting to note, however, that the composition of the normative feldspar and the percentage of iron in the femic constituents are nearly constant and

in reasonable agreement with the actual mineral compositions (see tables I, II, and III).

Comparison can be made with three analysed 'gabbroic' xenoliths from *Tristan da Cunha* (Dunne, 1941, table 14), all of which, however, are notably higher in TiO_2 , Fe_2O_3 , and CaO and lower in SiO_2 than those from Gough Island. Their norms (Baker *et al.*, 1964) show them to be highly undersaturated, no doubt due to the presence of the kaersutitic amphibole that is an essential characteristic of these xenoliths. From a preliminary examination of the numerous xenoliths collected by the Royal Society Expedition to Tristan da Cunha, 1962, it seems probable that these too are a heterogeneous mixture of mineral phases of more or less constant composition (a view held by Dunne), although the picture is somewhat complicated by reactions involving both the pyroxene and amphibole. The compositions of the mineral phases are, of course, different from those on Gough Island, kaersutitic amphibole being abundant while olivine is extremely rare and hypersthene absent, even as exsolution lamellae in the augite.¹ No peridotitic xenoliths have yet been found on Tristan.

Gabbroic xenoliths have also been recorded from many other volcanic provinces often in subordinate amounts to peridotitic xenoliths. Among other oceanic occurrences are:

Ascension Island where granitic, dioritic, and syenitic xenoliths have been recorded as well as the more usual gabbroic and peridotitic types, which sometimes contain hypersthene. The plagioclase in the gabbroic types has a composition of An_{75} (Daly, 1925). A series of specimens collected by Professor D. S. Coombs from Ascension and now in the Harker Collection, Cambridge University (69498-69560, 71591-71736), includes many basic xenoliths. These have been examined and have many features in common with those from Gough Island, including high modal variability, exsolution of hypersthene in augite and vice versa, and exsolution of pale yellow-green spinel in the pyroxene.

The *Kerguelen* Archipelago where both gabbroic (plagioclase An_{65}) and peridotitic types occur (Edwards, 1938; Talbot *et al.*, 1963). Hypersthene is present and both pyroxenes exhibit exsolution. Biotite and basaltic hornblende-bearing xenoliths occur in some of the basic alkali lavas, often with abundant apatite. This association of amphibole and apatite also occurs on Tristan. (Baker *et al.*, 1964).

Hawaii, where gabbroic and peridotitic types, which grade into each other, occur abundantly in the alkali lavas (Macdonald, 1949) but are

¹ This has been confirmed by single crystal X-ray methods.

almost absent in the tholeiitic lavas (Powers, 1955). There is a marked similarity between the gabbroic types from Gough and those from the 1801 lava, north of Hualalai (Harker Collection, 81862–81878); the olivines contain stringers of iron-ore and show signs of strain; augite exsolves hypersthene and vice versa; and both pyroxenes exsolve iron-ore and a transparent spinel.

Fayal, in the Azores, where dunites, gabbros (with and without olivine), and pyroxenites occur in that order of abundance (personal communication with M. J. Baker). No amphiboles or orthopyroxenes have so far been observed.

There are certain consistent features displayed by these gabbroic xenoliths that should be explained in any theory accounting for their origin: they have broad chemical affinities with the lavas in which they occur and are usually more basic than the erupted basaltic magma; hypersthene is of common occurrence especially in those types devoid of amphibole; the clinopyroxenes lie somewhere between typically tholeiitic and typically alkali types; they appear to represent fragments of a modally heterogeneous body composed of mineral phases of approximately constant composition; and their rarity in tholeiitic lavas compared with their abundance in alkali lavas.

An excellent review of the origin of gabbroic and also peridotitic xenoliths has been given by Wilshire and Binns (1961), who concluded that the xenoliths are derived from the magma source area in the mantle. These conclusions were based mainly on a 'deduction by elimination' method and by speculation on high-pressure mineralogy. However, now that a considerable amount of experimental work has been done on natural basalts and eclogites (Yoder and Tilley, 1962), many more observations are available concerning high-pressure phase relations and magma production, none of which detract from the 'origin in the mantle' hypothesis. In the light of these facts, therefore, the following slightly more specific hypothesis is proposed: that the xenoliths are derived in the mantle from depths of 40 to 50 km below the surface¹ and that the bulk composition of the mantle in this region is essentially that of an olivine-tholeiite as defined by Yoder and Tilley (p. 352).

Seismic evidence suggests that magma is produced 50 to 60 km beneath the Klyuchevsky volcano, Kamchatka (Gorskov, 1956), and from

¹ In terms of the pressure-temperature diagram for melts of basaltic compositions (Yoder and Tilley, 1962, p. 498, fig. 43) this places them, at liquidus temperatures, just above the transformation zone. If a shallower depth is postulated it is difficult to explain their rarity in tholeiitic magmas, while if a greater depth is postulated it is difficult to explain the rarity of eclogite xenoliths.

45 to 60 km beneath Kilauea, Hawaii (Eaton and Murata, 1960). This places these magma-producing areas in the transformation zone. Partial melting of material of essentially olivine-tholeiite composition at this depth would, under these relatively low-pressure conditions, yield a tholeiitic magma (Yoder and Tilley, 1962, p. 509), i.e. the type of magma produced at Kilauea and possibly Klyuchevsky (? calc-alkali magma = tholeiite + contamination by sial?), in which the xenoliths would be rare owing to the proximity of the magma source to the zone from which the xenoliths are thought to come.¹ At considerably greater depths, partial melting of material (eclogite?) of olivine-tholeiite composition would produce an alkali magma (Yoder and Tilley, 1962, p. 509) and could easily incorporate xenoliths as it passed through the cooler upper mantle. This could explain one of the most embarrassing problems of the xenoliths, their rarity in tholeiitic lavas. Confirmation of these ideas will, however, have to wait until there is seismic evidence for the depth of production of an alkali magma.

There are several reasons for postulating that the composition of the upper mantle is essentially olivine-tholeiite: the composition of the true gabbroic and peridotitic xenoliths, i.e. the olivine-clinopyroxene-orthopyroxene-plagioclase types, is essentially that of an olivine-tholeiite; if the two extreme types of fundamental magma (alkali and tholeiitic) are to be produced from a common source (Yoder and Tilley, 1962, p. 509) then the most reasonable composition for that source would be one intermediate between the two extremes i.e. olivine-tholeiite; in the system $Ne-Fo-SiO_2$ (Yoder and Tilley, 1962, p. 508) one of the regions from which either an alkali-rich or silica-rich liquid can be produced (depending on pressure) has the approximate composition of olivine-tholeiite; the hypersthene-eclogite from Salt Lake Crater, Oahu (Yoder and Tilley, 1962, p. 482), which, of the eclogites studied by Yoder and Tilley, has the best claim to genetic relation with basalt magma, is of olivine-tholeiite composition; and the pyroxenes might reasonably be expected to come from a basalt intermediate to alkali and tholeiitic types (olivine tholeiite?) as they exhibit characteristics of both these magma types.

As it has been observed that the xenoliths are modally heterogeneous it must also be assumed that the mantle is modally heterogeneous, at

¹ From their somewhat angular nature, often friable texture, and genuine plutonic appearance it is felt that they are brecciated fragments from a relatively brittle rock mass. This and the absence of large-scale melting effects would seem to preclude an origin near the magma source where the general increase in temperature would cause the rocks to become fairly plastic.

least on a small scale. This would seem to be a reasonable supposition as it is highly likely that the mantle has been subjected to some primitive differentiation process during its formation, which could lead to local layering, etc. However, this small-scale heterogeneity will not seriously affect the composition of magmas produced by partial melting.¹ This is because the composition of liquids derived by partial melting follows a unique path along the equilibrium boundary 'curves' (Yoder and Tilley, 1962, p. 397), which are defined not by the bulk composition but by the composition of the mineral phases involved and these, as we have seen, are nearly constant at any given locality.

Any process of primitive differentiation in the mantle would almost certainly preserve any local chemical peculiarities throughout the 'differentiates'. Thus xenoliths (= mantle), incorporated in a magma derived by partial melting from mantle material at a greater depth, would have chemical affinities with the lavas in which they occur, and because of the partial melting would also tend to be more basic. Furthermore, from theoretical considerations the partial crystallization of a magma produced by partial melting will precipitate several, but not necessarily all, of the phases present in the parent rock. Thus the fact that xenolithic minerals are often similar to the phenocrysts of the lavas in which they occur need not, as has often been assumed in the past (see review of Wilshire and Binns, 1961), indicate an origin by the accumulation of phenocrysts.

Regarding the amphibole-bearing xenoliths, the author thinks that these represent normal gabbroic xenoliths that have had time to react thoroughly with a hydrous magma at a relatively shallow depth in the crust. This is supported by experimental work on the conversion of basalt to hornblende-gabbro under hydrous conditions (Yoder and Tilley, 1962, p. 430), and by the absence of hypersthene as discrete crystals or as exsolution lamellae in the clinopyroxene since hypersthene reacts readily with alkali basalt magma to produce olivine.

In conclusion the general picture would not appear to be greatly affected by the type of material at depth, e.g. eclogite or peridotite, as long as the partial melting of whatever material is present will give an alkali basalt magma at a greater depth than a tholeiitic magma. The author considers the evidence to be in favour of eclogite, since if there

¹ Partial melting or crystallization is now considered to be essential to any process of basalt magma production (Yoder and Tilley, 1962, p. 397). Of these two the author considers partial melting to be more likely, mainly on the grounds that, as it would require a smaller rise in temperature, it would be likely to occur more often.

is no eclogite layer it is extremely difficult to explain the presence of the, admittedly rare, eclogite xenoliths. Peridotitic xenoliths, however, can always be derived by some primitive 'differentiation' process in the upper 'basaltic' mantle.

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References

- BAKER (P. E.), GASS (I. G.), HARRIS (P. G.), and LE MAITRE (R. W.), 1964. Volcanological report of the Royal Society Expedition to Tristan da Cunha, 1962. Phil. Trans. Roy. Soc. (A), vol. 256, p. 439.
- BOWN (M. G.) and GAY (P.), 1959. Amer. Min., vol. 44, p. 592.
- DALY (R. A.), 1925. Proc. Amer. Acad. Arts Sci., vol. 60, no. 1.
- DUNNE (J. C.), 1941. Results of Norwegian Sci. Exp. to Tristan da Cunha, 1937-38, No. 2. Det norske videnskaps-akad. Oslo.
- EATON (J. P.) and MURATA (K. J.), 1960. Science, vol. 132, p. 925.
- EDWARDS (A. B.), 1938. B.A.N.Z. Antarc. Research Exp. 1929-31. Reports Series A, vol. 2, part 5, p. 72.
- GAY (P.) and LE MAITRE (R. W.), 1961. Amer. Min. vol. 46, p. 92.
- GORSKOV (G. S.), 1956. C.R. Acad. Sci. U.R.S.S., vol. 106, no. 4, p. 703.
- HENRY (N. F. M.), 1942. Min. Mag., vol. 26, p. 179.
- KUSHIRO (I.), 1960. Amer. Journ. Sci., vol. 258, p. 548.
- LE BAS (M. J.), 1962. *Ibid.*, vol. 260, p. 267.
- LE MAITRE (R. W.), 1962. Bull. Geol. Soc. Amer., vol. 73, p. 1309.
- MACDONALD (G. A.), 1949. *Ibid.*, vol. 60, p. 1541.
- POWERS (H. A.), 1955. Geochim. et Cosmochim. Acta, vol. 7, p. 77.
- SEGNIT (E. R.), 1953. Min. Mag., vol. 30, p. 218.
- SMITH (J. V.) and GAY (P.), 1958. *Ibid.*, vol. 31, p. 744.
- TALBOT (J. L.), HOBBS (B. E.), WILSHIRE (H. G.), and SWEATMAN (T. R.), 1963. Amer. Min., vol. 48, p. 150.
- VERMAAS (F. H. S.) and SCHMIDT (E. R.), 1959. Beit. zur Min. und Petr., Bd. 6, p. 219.
- WILSHIRE (H. G.) and BINNS (R. A.), 1961. Journ. Petrology, vol. 2, p. 185.
- YODER (H. S.) and TILLEY (C. E.), 1962. *Ibid.*, vol. 3, p. 342.
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