

Kaersutite-bearing plutonic xenoliths from Tristan da Cunha, South Atlantic

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SUMMARY. The xenoliths, which are gabbroic in character, consist of kaersutite, titanaugite, plagioclase, and iron ore in variable amounts. Biotite and olivine are also occasionally present. New analyses and trace element data are given for two xenoliths and their coexisting kaersutite, titanaugite, and plagioclase. After discussing several possible origins with the aid of a principal latent vector variation diagram, it is concluded that they represent the accumulation of phases that crystallized from the Tristan magma at depths between 5 and 25 km. Evidence is also presented to show that the parental Tristan magma may be of trachybasalt composition.

THE first petrographic descriptions of the plutonic xenoliths from Tristan were given by Campbell Smith (1930) and Dunne (1941), who also presented some chemical data. In 1962, the Royal Society Expedition collected nearly 250 plutonic xenoliths, which are now housed in the Department of Mineralogy, British Museum (Natural History) under the number BM 1962, 128. However, as they were only briefly investigated (Baker *et al.*, 1964), two were selected for more detailed work, the results of which are presented in this paper. The two specimens are 109, from a red tuff, 50 ft above sea level on the north side of Eastend of Sandy Point Gulch, and 248, from a lava flow 40 ft below the cattle crossing in Gypsy's Gulch. Both types of environment in which the xenoliths can be found (i.e. lava flows and pyroclastic deposits) are thus represented.

With one notable exception, all the plutonic xenoliths described from Tristan are gabbroic in texture, and consist essentially of kaersutite, titanaugite, plagioclase, and iron ore in variable amounts. Olivine and biotite are also occasionally present. The exception is a water-worn block of gneiss picked up near the settlement of Edinburgh (Schwarz, 1905), which is almost identical to a specimen (No. 686) collected by the Royal Society Expedition from the beach at Blenden Hall on Inaccessible Is., the site of a notorious shipwreck (Mackay, 1963). In view of this, and the islanders habit of collecting things, it is felt that no petrological significance can be attached to these finds. It is unfortunate that evidence such as this has been perpetuated in the literature as evidence of sialic crust in the Tristan area (Gilluly, 1955).

Of particular interest is the fact that, to the best of the author's knowledge, none of the peridotitic or ultrabasic xenoliths so common in many of the alkali basalt provinces (e.g. Forbes and Kuno, 1965; White, 1966) have yet been found.

Mineralogy

Kaersutite. The chemical analyses and optics of the kaersutites from xenoliths 109 and 248 are given in table I together with K, a xenolithic amphibole analysis taken

from Dunne (1941, p. 50, Anal. 1). The optics of 109 and 248 were determined partly by measurement and partly by calculation. By measuring α' and γ' on (110) cleavage flakes, $2V_\alpha$, and $\gamma:[001]$ it was possible to calculate α and β , assuming that γ is only

TABLE I. *Chemical analyses, structural formulae, and optics of kaersutites*

	109	248	K	572	Number of ions on basis of 24 (O, OH, F, Cl)				
SiO ₂	37.4	37.6	39.95	37.70					
TiO ₂	6.0	7.1	8.79	6.42		109	248		
Al ₂ O ₃	15.7	13.0	9.81	12.35			K		
Fe ₂ O ₃	12.4	7.7	9.19	—	Z	Si	5.568	5.654	5.864
FeO	1.1	3.7	—	12.55*		Al ^{IV}	2.432	2.299	1.697
MnO	0.2	—	0.08	0.21		Ti ^{IV}	—	0.047	0.439
MgO	11.4	13.6	12.95	11.12	Y	Al ^{VI}	0.323	—	—
CaO	11.9	10.6	13.22	13.48		Ti ^{VI}	0.672	0.758	0.531
Na ₂ O	2.3	3.6	2.94	2.00		Fe ³⁺	1.389	0.871	1.015
K ₂ O	1.5	1.8	1.54	n.d.	X	Fe ²⁺	0.137	0.467	—
H ₂ O+	0.1	0.5	1.59	n.d.		Mn	0.025	—	0.010
H ₂ O-	0.1	—	—	n.d.		Mg	2.530	3.053	2.833
F	0.4	0.6	n.d.	n.d.	W	Ca	1.899	1.699	2.079
Cl	0.9	0.1	n.d.	n.d.		Na	0.664	1.057	0.837
Total	101.4	99.9	100.06	95.83		K	0.285	0.351	0.288
Less O	0.4	0.3	—	—	OH	0.070	0.491	1.557	
Total	101.0	99.6	100.06	95.83	F	0.179	0.289	—	
					Cl	0.227	0.031	—	
α' on 110†	1.722	1.704	—	—	W				
γ' on 110†	1.756	1.724	—	—		X	2.848	3.107	3.204
					Y				
α ‡	1.694	1.682	—	—		X	2.692	3.520	2.843
β ‡	1.730	1.710	—	—	Y	2.384	1.629	1.546	
γ §	1.757	1.725	—	—	Z	8.000	8.000	8.000	
$\gamma-\alpha$ ‡	0.063	0.043	—	—	WXY	7.924	8.256	7.593	
				100 Mg/(Mg+Fe ³⁺ +Fe ²⁺ +Mn)		61.9	69.5	73.5	
$2V_\alpha$ †	80°	70°	—	—					
$\gamma:[001]$ †	4°	2°	—	—					

* Total iron as FeO. † Measured value. ‡ Calculated value. § Assumed value.

572: From trachyandesite 572 (Baker *et al.*, 1964, p. 499). Mean of four partial electron-probe anal. by A. T. Anderson.

109: From xenolith in tuff, near Sandy Point. Anal. C. J. Elliott.

248: From xenolith in lava, Gypsy's Gulch. Anal. A. J. Easton.

K: From xenolith, locality unspecified (Dunne, 1941, p. 50, Anal. 1). Anal. J. Jakob.

just greater than γ' (their positions on the indicatrix differ by only a few degrees) and (110):(110) = 56°. Baker *et al.* (1964), for want of more precise data, called these strongly pleochroic red-brown amphiboles basaltic hornblende.

Aoki (1963) contends that the degree of oxidation of kaersutite is controlled by the mechanism of eruption, the process being achieved mainly by the (OH,F) in the amphibole, by a double replacement of the type $\text{Fe}^{2+}(\text{OH},\text{F})^{1-} = \text{Fe}^{3+}\text{O}^{2-}$. This is confirmed by the present study as the kaersutite 109 ($\text{Fe}^{3+}/\text{Fe}^{2+} = 8.6$ and OH,F,Cl

= 0.476) comes from a xenolith in a red scoriaceous tuff (highly oxidizing environment) while the kaersutite 248 ($\text{Fe}^{3+}/\text{Fe}^{2+} = 1.9$ and $\text{OH,F,Cl} = 0.811$) is from a xenolith in a lava (less oxidizing environment). Furthermore, Aoki notes that kaersutites from monzonite and teschenite, unlike those from volcanic rocks, often have insufficient Al^{3+} to fill the Z group. If this is a valid observation both 248 and K may not have originally crystallized under volcanic conditions.

The only data available for comparison with the phenocrystic amphiboles in the Tristan lavas are some partial electron-microprobe analyses, kindly made available by Dr. A. T. Anderson of the United States Geological Survey, of the amphibole in the trachyandesite 572. Baker *et al.* (1964) describe the amphibole as being apparently in equilibrium and coexisting with phenocrysts of mildly resorbed biotite and zoned plagioclase. Although they suggest it may be a barkevikitic hornblende, it is obviously a kaersutite similar in chemistry and optics to 248 (see table I).

Titanaugite. The chemical analyses and optics of the titanaugites from xenoliths 109 and 248 are given in table II. In addition four analyses are given of pyroxene phenocrysts from Tristan lavas arranged in order of increasing differentiation. Analyses B, C, and D are the averages of several electron microprobe analyses, also made available by Dr. A. T. Anderson of the United States Geological Survey, of pyroxenes from rocks described by Baker *et al.* (1964).

In discussing the crystallization trends of titanaugites Yagi and Onuma (1967) point out that during the early stages of differentiation the titanaugites are progressively enriched in Ti and that in many provinces this trend is reversed in the middle and late stages of differentiation to give Ti-poor pyroxenes in the late differentiates. This trend is poorly shown by the titanaugites A, B, C, and D with a weak maximum at C. However, as no examples are known of Ti impoverishment throughout the whole of a differentiation series, the titanaugites from the xenoliths 109 and 248 cannot be early formed phenocrysts as their Ti content is greater than any of the phenocrysts. Likewise their Na content is also too high and their Ca:Mg:Fe ratio wrong. Therefore, for the titanaugites 109 and 248 to be phenocrysts they would have to fit towards the end of the differentiation series somewhere between C and D. However, in this position their Al contents are too high and their Si contents too low. The titanaugites 109 and 248 do not, therefore, appear to belong to the same generation as the phenocrystic titanaugites.

Furthermore, the titanaugites 109 and 248 have higher Al_2O_3 contents than any of the phenocrystic titanaugites suggesting that the xenolithic titanaugites crystallized under higher pressure conditions than the phenocrysts (Kushiro, 1965). However, a very approximate upper limit of 10 kb. can be placed on this higher pressure regime as the experiments of Yagi and Onuma (1967) indicate that above this pressure the solubility of $\text{CaTiAl}_2\text{O}_6$ in diopside decreases to nil.

Unlike the clinopyroxenes in the gabbroic xenoliths from Gough Island (Le Maitre, 1965) none of the xenolithic clinopyroxenes from Tristan contains any signs of exsolution of other pyroxene phases when examined by single crystal X-ray methods as advocated by Bown and Gay (1959). This also would seem to indicate an origin not too deep in the crust.

TABLE II. *Chemical analyses, structural formulae, and optics of titanaugites*

	109	248	A	B	C	D	
SiO ₂	45.0	43.8	48.70	45.56	46.41	47.75	
TiO ₂	3.1	3.2	2.42	2.22	2.63	2.16	
Al ₂ O ₃	9.1	8.8	4.69	7.94	6.54	4.98	
Fe ₂ O ₃	4.5	5.0	2.30	8.40*	7.67*	9.87*	
FeO	4.6	3.7	5.00				
MnO	0.4	0.1	0.12	0.10	0.12	0.37	
MgO	11.2	12.0	14.38	12.55	12.64	10.81	
CaO	21.9	22.9	21.89	22.78	23.54	23.30	
Na ₂ O	0.7	0.7	0.64	0.45	0.45	0.76	
K ₂ O	< 0.1	< 0.1	0.06	n.d.	n.d.	n.d.	
H ₂ O	—	—	0.04	n.d.	n.d.	n.d.	
	100.5	100.2	100.24	100.00	100.00	100.00	
<i>Number of ions on basis of 6 oxygens</i>							
Z	Si	1.678	1.644	1.806	1.718	1.748	1.814
	Al ^{iv}	0.322	0.356	0.194	0.282	0.252	0.186
Al ^{vi}	Al ^{vi}	0.078	0.033	0.011	0.071	0.038	0.037
	Ti	0.087	0.090	0.068	0.063	0.074	0.062
Fe ³⁺	Fe ³⁺	0.126	0.141	0.064	0.265*	0.242*	0.314*
	Fe ²⁺	0.143	0.116	0.155			
X	Mn	0.013	0.003	0.004	0.003	0.004	0.012
	Mg	0.623	0.671	0.795	0.705	0.709	0.612
Ca	Ca	0.875	0.921	0.870	0.920	0.950	0.949
	Na	0.051	0.051	0.046	0.033	0.033	0.056
K	K	0.000	0.000	0.003	—	—	—
	Z	2.000	2.000	2.000	2.000	2.000	2.000
X	1.996	2.026	2.016	2.060	2.050	2.042	
Ca	49.2	49.8	46.1	48.7	49.9	50.3	
Mg	35.0	36.2	42.1	37.2	37.2	32.4	
Fe	15.8	14.0	11.8	14.1	12.9	17.3	
β	1.713	1.721	1.705	—	1.717	1.708	
2V _γ	48°	49°	56°	—	50°	59°	

* Total iron determined as FeO.

109: Titanaugite. Xenolith in tuff, near Sandy Point. Anal. C. J. Elliott.

248: Titanaugite. Xenolith in lava, Gypsy's Gulch. Anal. A. J. Easton.

A: 'Basaltic augite.' Olivine basalt, Sandy Point (Dunne, 1941, p. 44, Anal. 1). Anal. J. C. Dunne.

B: 'Pyroxene.' Trachybasalt 622, Settlement. Average of four electron microprobe analyses. Anal. A. T. Anderson.

C: 'Titaniferous augite.' Trachybasalt 347, Pigbite Gulch. Average of five electron microprobe analyses. Anal. A. T. Anderson.

D: 'Aegirine-augite.' Plagioclase trachyte 560. The Peak. Average of seven electron microprobe analyses. Anal. A. T. Anderson.

Plagioclase. Partial chemical analyses of the plagioclases from two xenoliths gave, for nos. 109 and 248 respectively: CaO 15.5, 14.8; Na₂O 1.9, 1.5; K₂O 0.1, 0.1%; Or:Ab:An, 1:17:82, 1:14:85 (wt %); Σ (Or, Ab, An) 94, 87%. No. 109 is from a tuff near Sandy Point, anal. C. J. Elliott, and no. 248 from lava at Gypsy's Gulch, anal. J. Easton. Recalculated in terms of the feldspar end-members these give

compositions even more basic than those suggested by Campbell Smith (1930), Dunne (1941), and Baker *et al.* (1964) from optical studies. The only plagioclases occurring in the lavas as basic as these are in the cores of some of the phenocrysts, e.g. ankaramite 114, trachybasalts 347 and 351 (Baker *et al.*, 1964). Therefore, if the plagioclases 109 and 248 are related to the phenocrysts they must have crystallized at a very early stage.

Petrography

The petrography is confined to xenoliths 109 and 248 as general descriptions of the xenoliths can be found in Campbell Smith (1930), Dunne (1941), and Baker *et al.* (1964).

Xenolith 109, from a red tuff, is a friable rock in hand specimen with abundant roughly orientated blades of black-brown glistening kaersutite up to 10 mm long and 2 mm wide, black granular titanaugite and slightly iron-stained interstitial plagioclase up to 3 mm across. In thin section the texture is granular to gabbroic. The subhedral blades of *kaersutite* are extremely fresh in appearance and are strongly pleochroic from α pale straw to γ dark red-brown. They show little signs of zoning and contain a few inclusions of granular iron ore and rarer titanaugite and plagioclase. The pale grey-green *titanaugite* is mainly anhedral and shows slight zoning. A few simple twins are present, as are inclusions of iron ore, kaersutite, and plagioclase. Small irregular patches of optically orientated kaersutite, which are distinct from the inclusions, are occasionally developed in the titanaugite. The *plagioclase*, which is fairly fresh and relatively unzoned, occurs interstitially as small multiple-twinned interlocking laths up to 2 mm long with a few inclusions of iron ore. Apart from inclusions, *iron ore* also occurs as both discrete equidimensional grains up to 1 mm across and skeletal masses enclosing the plagioclase laths.

Xenolith 248, from a lava flow, is a more compact rock in hand specimen than 109 and consists of granular black kaersutite and titanaugite with interstitial plagioclase. In thin section the texture is also granular to gabbroic. The *kaersutite*, which is pleochroic from α pale straw to γ dark brown, tends to occur as laths up to 4 mm long. It contains a few inclusions of iron ore and plagioclase and more rarely titanaugite. In many places around the margins of the crystals and along cracks it is breaking down to a fine granular mosaic of iron ore, titanaugite, plagioclase, and a pleochroic, red-brown to green-brown, platy mineral of low birefringence and high relief, which is probably the mineral referred to by Dunne (1941) as rhönite, although the habit is different. A similarly coloured mineral occurring in a crystalline aggregate around kaersutite has been described by Uchimizu (1966) as an alkali amphibole. The *titanaugite*, which occurs as anhedral crystals up to 2 mm across, is pale grey and only slightly zoned. The only inclusions are a few small rounded grains of iron ore. Irregular patches of kaersutite in optical continuity are more abundant than in 109 and many of the crystals contain minute orientated rods of an obviously exsolved phase (ilmenite?, spinel?), which except for colour are similar to those described in the gabbroic xenoliths from Gough Island (Le Maitre, 1965). The *plagioclase* occurs as rounded interstitial laths up to 1 mm long, with a few inclusions of rounded blebs

of pyroxene and minor ore. The *iron ore* occurs as irregular grains up to 1 mm across that are interstitial to the kaersutite and titanaugite. It is, however, often separated from the titanaugite by a thin zone of plagioclase crystals up to 0.05 mm wide.

Reaction in the xenoliths

Two reactions are commonly found within the xenoliths. One is the conversion of titanaugite to kaersutite, which ranges from submicroscopic intergrowths only

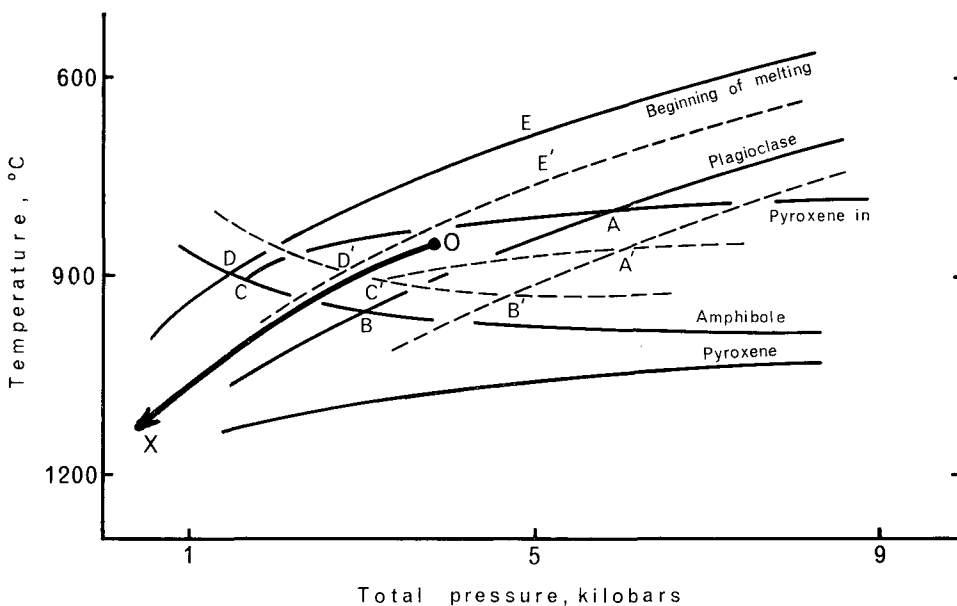


FIG. 1. Hypothetical P - T diagram for Tristan magma under different conditions of water pressure. Solid lines—water pressure equals total pressure (based on the oxidised hawaiiite of Yoder and Tilley (1962, fig. 20)). Stability field of amphibole+pyroxene+plagioclase, ABC ; amphibole+plagioclase, $ACDE$. Dashed lines—water pressure less than total pressure (deduced from Yoder and Tilley (1962, fig. 34)). Stability field of amphibole+pyroxene+plagioclase, $A'B'C'$; amphibole+plagioclase, $A'C'D'E'$. OX —possible P - T course for rise of xenolith to surface.

detectable by X-ray methods to large optically orientated patches as described in 109 and 248. This reaction is well developed in most of the xenoliths from pyroclastic centres and to a lesser extent in those from lava flows. As the stable assemblage in the xenoliths appears to be amphibole+pyroxene+plagioclase this means that, before eruption, the xenoliths must move into the stability field of amphibole+plagioclase. The hypothetical position of these stability fields under different conditions of water pressure are shown in fig. 1. This is based on the data of Yoder and Tilley (1962, fig. 30 and fig. 34). The oxidized hawaiiite was chosen for its similarity to the Tristan trachybasalts in chemistry and melting relationships (Tilley *et al.*, 1965). Similar but smaller stability fields also exist at lower pressures for the alkali basalt of Yoder and

Tilley (1962, fig. 20). If, as seems likely (see p. 196), the xenoliths are formed by the accumulation of phases crystallizing at depth the reaction could occur as a normal consequence of falling temperature. It could also occur if the ratio of H_2O pressure to total pressure decreases. Once the xenolith starts moving towards the surface along some such path as OX in fig. 1 (see Harris, 1962) the only reasonable way in which it could enter the amphibole+plagioclase field would again be by a reduction in the ratio of H_2O pressure to total pressure in the melt. This could be achieved by the development of a discrete gas phase resulting from a reduction in total pressure as described by Yoder (1965), which would also explain why the reaction is more prominent in pyroclastic eruptions.

The other reaction is the breakdown of kaersutite to a granular mosaic of minerals that have been described, here and elsewhere (Dunne, 1941; Baker *et al.*, 1964) as including plagioclase, titanite, iron ore, rhönite, olivine, biotite, ilmenite, and alkali amphibole. This is almost certainly a near surface effect reflecting the fact that amphibole is not a stable phase even in hydrous basaltic melts at water pressures less than 1.4 kb (Yoder and Tilley, 1962). The fact that this reaction is not found in xenoliths from pyroclastic centres is attributed to the relatively rapid chilling of the ejecta during such an eruption. This effect can be seen in the 1961 trachyandesite eruption where kaersutite phenocrysts in the crystalline lavas are frequently resorbed while those from glassy bombs are not (Baker *et al.*, 1964).

Chemistry of the xenoliths

The chemical analyses and C.I.P.W. norms of xenoliths 109 and 248 are given in table III together with those of three xenoliths investigated by Dunne (1941), the norms of which have not previously been published. The most characteristic features of these analyses are their high TiO_2 contents and Fe_2O_3/FeO ratios coupled with a degree of SiO_2 undersaturation that, considering the Fe_2O_3/FeO ratio, is remarkably high. They are, therefore, completely unlike the gabbroic xenoliths from neighbouring Gough Island (Le Maitre, 1965).

Trace element data for xenoliths 109 and 248 and some of their individual mineral phases are given in table IV. Comparative data for the distribution of elements in coexisting kaersutite and titanite appears to be lacking. However, White (1966) in a study of the ultrabasic inclusions from Hawaii comments on the distribution of several elements between coexisting hornblende and clinopyroxene. He notes that Ti, Cr, Ni, Na, and K are more abundant in the hornblende than in the clinopyroxene while the reverse is true for Mn. These distributions are the same as those between the coexisting kaersutites and titanites, although in the case of Cr and Ni the data are poor. Nb is markedly concentrated in the kaersutites, which agrees with the data of Huckenholz (1965), although the absolute amounts present are less by at least a factor of 2 to 3. From the distribution of Cr, V, Ni, Co, and Pb between the xenoliths and their analysed minerals, it is also obvious that these elements are enriched in the iron-ore phases, which have not been analysed. The low Cr contents of the xenoliths are also worthy of note, being less than those of the basalts and more comparable to those of the trachybasalts (Baker *et al.*, 1964).

TABLE III. *Chemical analyses and C.I.P.W. norms of xenoliths*

	109	248	AX	HX	BX
SiO ₂	34.9	37.5	35.70	36.76	40.76
TiO ₂	5.5	4.6	5.13	4.88	3.56
Al ₂ O ₃	17.8	16.0	10.45	16.18	19.19
Fe ₂ O ₃	15.7	11.1	10.88	13.13	7.62
FeO	1.5	4.3	9.54	3.87	2.29
MnO	0.3	0.1	0.12	0.06	0.03
MgO	9.0	7.7	9.75	7.50	6.05
CaO	12.9	15.9	16.32	14.98	14.07
Na ₂ O	1.9	1.5	1.53	1.59	2.42
K ₂ O	0.8	0.5	0.14	0.33	1.36
P ₂ O ₅	n.d.	< 0.1	0.05	0.02	1.99
H ₂ O+	< 0.1	0.1	0.61	0.83	0.70
H ₂ O-	< 0.1	< 0.1	—	—	—
F	0.3	0.2	n.d.	n.d.	n.d.
Cl	n.d.	< 0.1	n.d.	n.d.	n.d.
	100.5	99.4	100.22	100.13	100.04
<i>C.I.P.W. norms</i>					
Or	4.73	—	—	—	8.04
Ab	3.19	—	—	—	10.92
An	37.91	35.45	21.24	36.04	37.49
Lc	—	2.32	0.65	1.53	—
Ne	6.75	6.88	7.01	7.29	5.17
Di	8.90	27.86	34.13	27.93	12.38
Ol	12.81	4.39	6.12	4.02	6.54
Cs	—	1.91	4.84	0.16	—
Mt	—	0.85	15.77	—	—
Hm	15.65	10.51	—	13.13	7.62
Il	3.81	8.74	9.74	8.30	4.90
Pf	5.95	—	—	0.87	1.67
HAp	—	—	0.12	0.05	4.69
Fl	0.70	0.41	—	—	—
H ₂ O	—	0.10	0.61	0.83	0.62
	100.40	99.42	100.22	100.13	100.04
Fe/(Fe+Mg)	0.00	0.00	0.01	0.00	0.00
<i>Modes</i>					
Plagioclase	27	36	11.6	32.4	52.2
Titanaugite	24	21	60.7	27.6	4.2
Kaersutite	43	27	7.9	27.8	24.6
Biotite	—	—	—	—	14.0
Ore etc.	6	16	19.8	12.2	6.0

109: Xenolith from tuff, near Sandy Point. Anal. C. J. Elliott.

248: Xenolith from lava, Gypsy's Gulch. Anal. J. Easton.

AX: 'Augitic endogenic inclusion', basalt, Sandy Point (Dunne, 1941, p. 93, Anal. 1). Anal. J. Jakob.

HX: 'Hornblenditic endogenic inclusion', basalt, Sandy Point (Dunne, 1941, p. 93, Anal. 2). Anal. J. Jakob.

BX: 'Biotite-bearing plagioclase-rich endogenic inclusion in red cinder', Half Way Beach (Dunne, 1941, p. 93, Anal. 3). Anal. J. Jakob.

In order to show the relationship between the xenoliths and the Tristan lavas use is made in fig. 2 of the principal latent vector variation diagram as described in detail by Le Maitre (1968).

TABLE IV. *Major and minor elements of the xenoliths and their minerals (anal. C. J. Elliott)*

	Xenolith		Kaersutite		Titanaugite		Plagioclase	
	109	248	109	248	109	248	109	248
<i>Major, %</i>								
Si	163	175	175	180	210	204	218	n.d.
Ti	33	28	36	38	19	19	1	n.d.
Al	94	85	83	65	48	47	180	n.d.
Fe ³⁺	110	78	87	50	31	35	7	n.d.
Fe ²⁺	12	33	9	29	36	29	n.d.	n.d.
Mg	54	46	68	80	67	72	2	n.d.
Ca	92	114	85	91	156	164	111	106
Na	14	11	17	23	5	5	14	11
K	7	4	12	12	1	tr.	1	1
ΣFe	122	111	96	79	67	64	7	—
<i>Minor, ppm</i>								
Nb	15	—	15	20	—	—	—	—
Zr	90	100	120	125	180	180	< 10	< 10
Ga	23	25	16	22	19	20	29	26
Cr	35	—	< 10	—	—	—	< 10	—
V	400	500	350	350	320	350	25	50
Y	20	—	20	35	35	40	—	—
Sc	40	40	50	70	100	120	< 20	< 20
Be	11	13	11	13	8	14	—	—
Ni	40	—	22	—	—	—	—	—
Co	70	70	35	60	22	27	—	—
Sr	50	85	30	n.d.	< 10	n.d.	50	70
Pb	10	8	4	3	3	5	4	< 3
Ga/Al*	0.24	0.29	0.19	0.34	0.40	0.43	0.16	—
V/Mg*	7.4	10.8	5.1	4.4	4.7	4.9	12.5	—
Ni/Mg*	0.74	—	0.32	—	—	—	—	—
Co/Mg*	1.3	1.5	0.51	0.75	0.33	0.38	—	—
Fe/Mg	2.3	2.4	1.4	0.99	1.0	0.89	3.5	—
V/Fe*	3.3	4.5	3.7	4.4	4.8	5.5	3.6	—
Y/Ca*	0.22	—	0.24	0.38	0.22	0.24	—	—
Sr/Ca*	0.54	0.75	0.35	—	< 0.06	—	0.45	0.66

* Ratio multiplied by 10³.

Briefly any chemical analysis of n oxides can be considered as a single point in n -dimensional space defined by n orthogonal oxide axes. By means of principal component analysis it is found that three orthogonal directions in n -space (the first three principal latent vectors) can account for over 90 % of all the chemical variation in volcanic rocks. The principal-latent-vector variation diagram is then the 2-dimensional representation of the projection of the analyses in n -space into the 3-space defined by the first three principal latent vectors. The great advantage of this type of diagram is

that it presents a very close approximation of the relative positions of the analyses in n -space in a simple diagram, unlike, for example, the silica-total-alkali diagram, which, although simple, contains no information about the other oxides. Obviously, if one is interested in the relationship of, say, three compositions A , B , and C , this

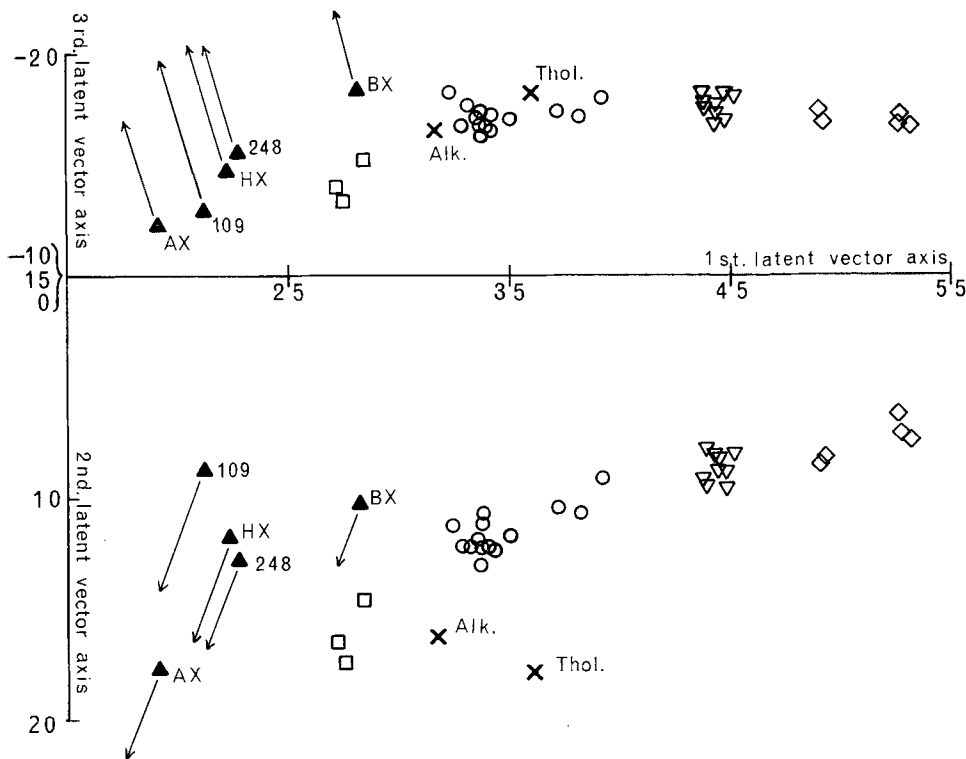


FIG. 2. Principal latent vector variation diagram of Tristan lavas and xenoliths using the same axes as in Le Maitre (1968). This figure is a 2-dimensional representation of a 3-dimensional projection in which the three latent vector axes are orthogonal. \square —ankaramite and olivine basalt; \circ —trachybasalt; ∇ —trachyandesite; \diamond —trachyte; \blacktriangle —xenolith (head of arrow indicates position to which xenolith would move if all iron were converted to FeO); \times —possible parental magmas.

is important. In this diagram, as in all linear projections, if $A+B=C$, the points representing these compositions must lie in a straight line. However, if $A+B \neq C$, the points may still lie in a straight line (in this case, only if the inequality is not too great), although as a corollary if the points do not lie on a straight line, $A+B \neq C$.

Fig. 2, which uses the same latent vector axes as those of Le Maitre (1968) shows that the xenoliths lie on a rough trend extending from the basic pyroxene-type AX, through the amphibole-types to the less basic biotite-type¹ BX—a compositional variation as great as that between the trachybasalts and trachyandesites. From the

¹ In this diagram the basic rocks plot to the left and the acid rocks to the right.

modes of the xenoliths given in table III it can be seen that variation along the trend is controlled mainly by the plagioclase content and across the trend by the amphibole/pyroxene ratio. However, although the trend is parallel to and extends beyond the basic end of the Tristan lava series it is clear that the xenoliths cannot be considered as a more basic extension of the lava series as the two trends are never coincident.

Origin of the xenoliths

Several of the possible hypotheses relating to the origin of these xenoliths will now be examined individually.

The xenoliths are completely accidental and have no genetic relationship with the lavas. This would seem to be unlikely as it has already been shown that there are many similarities between the mineralogy and chemistry of the lavas and xenoliths. White (1966) also tends to exclude this origin for the ultrabasic xenoliths occurring in the alkali basalts from Hawaii.

The xenoliths are completely accidental but have contaminated a 'normal' magma to produce the Tristan magma, which is more undersaturated and potassic than most oceanic alkali basalt series. For this hypothesis to be chemically valid the point representing the Tristan magma must lie on a straight line between the points representing the xenoliths and the 'normal' basalt in fig. 2. To represent possible 'normal' basalts both the averages of the alkali basalts (Alk.) and tholeiites (Thol.) of Poldervaart (1955) have been plotted. Although in general the Tristan magmas do lie between the xenoliths and the 'normal' basalts, the fit is rather poor. Even if allowance is made for minor variations in the positions of the 'normal' basalts to improve the fit, the xenoliths and the 'normal' basalts would still have to be added in almost equal amounts to produce a Tristan magma. In view of the fact that the 'normal' basalts are almost certainly products of partial melting, this raises the important problem of can such a magma assimilate so much material and still remain liquid; in the light of present knowledge the answer to this would appear to be no.

The xenoliths are Tristan magmas that have crystallized at depth. This is the hypothesis favoured by Baker *et al.* (1964) for the origin of the Tristan xenoliths. However, it has already been pointed out that in fig. 2 the xenoliths do not lie on the trend of the Tristan lava series or its basic continuation. The situation is not improved by assuming that most of the oxidation of the Fe in the xenoliths took place during the extrusive phases of volcanicity and that originally the xenoliths had a much lower $\text{Fe}_2\text{O}_3/\text{FeO}$ ratio, as this merely moves the xenoliths away from the trend of the Tristan lava series in the direction of the arrows, the points of which mark the positions of the xenoliths when all the Fe is expressed as FeO.

The xenoliths are the parent material from which the Tristan magma is derived by partial melting. Although it can always be postulated that the parent material for the generation of the Tristan magma is of unusual composition, this is most unsatisfactory as it merely pushes the problem of origin back a stage. Furthermore, as a mantle of this composition would almost certainly be unacceptable the region of Tristan magma generation would have to be above the mantle at a relatively shallow depth in the

crust, thus conflicting with recent theories on the origin of alkali magmas (O'Hara, 1965).

The xenoliths are the residue left when the parent material is partially melted to produce the Tristan magma. In this case the parent material must have a composition that plots between the position of the xenoliths and the Tristan magma in fig. 2. However, compositions close to those of the xenoliths have already been deemed unlikely in discussing the previous hypothesis. This only leaves compositions similar to the Tristan magma, which also means postulating a parent material of composition peculiar to Tristan. Generally such postulates are unacceptable. White (1966) also finds mantle hypotheses (i.e. this and the preceding hypothesis) unacceptable for the ultrabasic xenoliths found in the alkali basalts from Hawaii.

The xenoliths are accumulations of phenocrysts as found in the lavas. This is the hypothesis favoured by Dunne (1941) for the origin of these xenoliths. However, although the xenolithic kaersutite and plagioclase are similar to possible phenocrysts, the titanagites are not. Furthermore, olivine, which is an abundant phenocryst, is almost absent from the xenoliths, while kaersutite, which is an abundant mineral of the xenoliths, is not a common phenocryst. On the whole, therefore, this hypothesis seems improbable for the Tristan xenoliths.

The xenoliths are accumulations of phases crystallizing from the Tristan magma at depth. This is the hypothesis currently favoured by the author, partly by elimination and partly by the facts that the titanagites probably crystallized under pressures greater than those of the phenocrysts but less than approximately 10 kb (p. 187) and that amphibole+pyroxene+plagioclase is a stable assemblage in some hydrous basaltic melts at pressures greater than 1.4 kb and less than approximately 6 kb (fig. 1). This would place the depth of formation of the xenoliths somewhere between 5 and 25 km. White (1966) also favours a similar hypothesis for the origin of the ultrabasic xenoliths from the alkali basalts of Hawaii.

Furthermore, if it is assumed that the separation of these mineral phases from the magma is responsible for the differentiation of the lava series the more basic magma should lie on a straight line between the less basic magma and the xenoliths in n -space, and the phases present in the xenoliths should be capable of crystallizing from the more basic melt. Inspection of fig. 2 shows that for most of the lava series this is true, especially if it is also assumed, not unreasonably, that the $\text{Fe}_2\text{O}_3/\text{FeO}$ ratios of the xenoliths were originally lower. For example, the trachybasalt compositions lie on a straight line between the trachyandesites and xenoliths 248 and HX and the trachyandesites lie on a straight line between the trachytes and xenolith BX.

The change in mineral composition of the xenoliths from the kaersutite-types 248 and HX to the biotite-type BX, as crystallization proceeds from the trachybasalts to the trachyandesites, is what would be expected from a study of the lavas, as although kaersutite, titanagite, and plagioclase are present as phenocrysts in both the trachybasalts and the trachyandesites, biotite only appears in the latter (Baker *et al.*, 1964). It is also interesting to note that biotite is an important constituent in three out of four plutonic xenoliths described from the trachyandesite flows of the 1961 eruption (Baker *et al.*, 1964).

However, it is not possible to derive the trachybasalts from the olivine basalts in a similar manner, as the olivine basalts do not lie between the xenoliths and any of the other lavas. It is suggested, therefore, that the parent Tristan magma is a trachybasalt and that the olivine basalts are derived from it by the accumulation, at relatively high levels in the crust, of phenocrystic olivine and pyroxenes. This would also explain the observed frequency distribution of the lava types (Baker *et al.*, 1964) and the relatively low Cr and olivine content of the xenoliths. In the Tristan trachybasalts at 1 atm pressure olivine only occurs as a third liquidus phase after plagioclase and clinopyroxene (Tilley *et al.*, 1965).

The problem of the origin of the Tristan trachybasalts, however, still remains to be solved and is, unfortunately, beyond the scope of this paper.

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REFERENCES

- AOKI (K.), 1963. *Journ. Petrology*, **4**, 198–210.
 BAKER (P. E.), GASS (I. G.), HARRIS (P. G.), and LE MAITRE (R. W.), 1964. *Phil. Trans. ser. A.*, **256**, 439–578.
 BOWN (M. G.) and GAY (P.), 1959. *Amer. Min.*, **44**, 592–602.
 CAMPBELL SMITH (W.), 1930. In *Report of the 'Quest' Expedition*, Brit. Mus. (Nat. Hist.), chap. 8.
 DUNNE (J. C.), 1941. *Result of Norweg. Sci. Exp. to Tristan da Cunha, 1937–38*. Oslo (Norske Videnskaps-Akademi), **2**, 1–145.
 FORBES (R. B.) and KUNO (H.), 1965. In *The Upper Mantle Symposium, New Delhi, 1964*. Int. Union Geol. Sci. Copenhagen, 161–79.
 GILLULY (J.), 1955. *Bull. Geol. Soc. Amer.*, Special paper **62**, 7–18.
 HARRIS (P. G.), 1962. *Amer. Journ. Sci.* **260**, 783–6.
 HUCKENHOLZ (H. G.), 1965. *Geochimica Acta*, **29**, 807–20.
 KUSHIRO (L.), 1965. *Ann. Rep. Dir. Geophys. Lab. Washington, Year Book*, **64**, 112–17.
 LE MAITRE (R. W.), 1965. *Min. Mag.* **34**, 303–17.
 ——— 1968. *Journ. Petrology*, **9**, 220–52.
 MACKAY (M.), 1963. *Angry Island*. London (Barker).
 O'HARA (M. J.), 1965. *Scot. Journ. Geol.* **1**, 19–40.
 POLDERVAART (A.), 1955. *Bull. Geol. Soc. Amer.*, Special paper **62**, 119–44.
 SCHWARZ (E. H. L.), 1905. *Trans. South African Phil. Soc.* **16**, 9–51.
 TILLEY (C. E.), YODER (H. S.), and SCHAIRER (J. F.), 1965. *Ann. Rep. Dir. Geophys. Lab. Washington, Year Book*, **64**, 69–82.
 UCHIMIZU (M.), 1966. *Journ. Fac. Sci. Tokyo Univ.*, sec. 2, **16**, 85–159.
 WHITE (R. W.), 1966. *Beitr. Min. Petr.* **12**, 245–314.
 YAGI (K.) and ONUMA (K.), 1967. *Journ. Fac. Sci. Hokaido Univ.*, ser. 4, **13**, 463–83.
 YODER (H. S.), 1965. *Ann. Rep. Dir. Geophys. Lab. Washington, Year Book*, **64**, 82–9.
 ——— and TILLEY (C. E.), 1962. *Journ. Petrology*, **3**, 342–532.

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