MINERALOGICAL MAGAZINE, DECEMBER 1977, VOL. 41, PP. 501-11

Rubidium–strontium age studies and geochemistry of acid veins in the Freetown complex, Sierra Leone

R. D. BECKINSALE, J. F. W. BOWLES
Institute of Geological Sciences, Gray's Inn Road, London W1X 8NG

R. J. PANKHURST
British Antarctic Survey

AND

M. K. WELLS
University College, Gower Street, London WC1E 6BT

With chemical analyses by H. M. Blauer, University College, London WC1E 6BT

SUMMARY. The stratigraphical limits on the age of the Freetown intrusion, Sierra Leone, are very wide, yet the intrusion has not previously been accurately dated by isotopic methods, despite a number of attempts. Rubidium–strontium dating of acid veins contemporaneous with the early stages of the prolonged cooling history of the intrusion provides an age of 193 ± 3 Ma. The veins consist of quartz and orthoclase with relict minerals, principally plagioclase, from the host gabbro. Electron-microprobe analysis of the altered minerals of the veins, and the petrography of the vein and adjacent host gabbro clearly demonstrate that the veins were formed from a granitic fraction, differentiated *in situ* from the surrounding solid gabbro with the assistance of a hydrous fluid phase within the incipient vein. This assertion is supported by the identical, low value of the initial $^{87}Sr/^{86}Sr$ ratio (0·70389) obtained from both the acid and basic rocks, and the technique described here may be useful in dating other, similar, intrusions.

THE Freetown Complex is a large basic intrusion on the western seaboard of Sierra Leone and has been described in detail in a memoir by Wells (1962). Only the eastern part is exposed on land, as spectacular mountains rising to about 900 m above sea level. It consists of about 6000 m of rhythmically layered sequences of troctolitic, gabbroic, and anorthositic rocks, which show variation in their mineral proportions, but no progressive (cryptic) variation in the composition of the minerals. In part this may be due to the major rhythmic units of layering having been produced by separate episodes of intrusion. The Complex is intruded into Precambrian metamorphic rocks and overlain by sediments of the Bullom Group. The basal part of the Bullom Group is probably of Eocene Age. Thus stratigraphical limits on the age of the igneous intrusion are extremely wide. Knowledge of the age is important, however, with regard to possible connections between the intrusion of the coast-parallel dyke swarm of Liberia and that of Sierra Leone, the Freetown Complex, and the initial rifting that led to the formation of the Atlantic Ocean. Moreover, a magnetic polarity reversal occurred during the cooling of the Complex (Briden, Henthorn, and Rex, 1971), for which a reliable radiometric age would be important in helping to establish Mesozoic palaeomagnetic stratigraphy (McElhinny and Burek, 1971).

Previous attempts at dating the Freetown Complex have been confined to K–Ar methods and have proved difficult to interpret. Snelling (1966) published a summary of the K–Ar data determined at the Institute of Geological Sciences for plagioclase and pyroxene separates from the Freetown Complex provided by one of us (M. K. W.), which gave a range of conventional ages from 194 Ma to 1875 Ma. Snelling suggested that the oldest ages were probably more

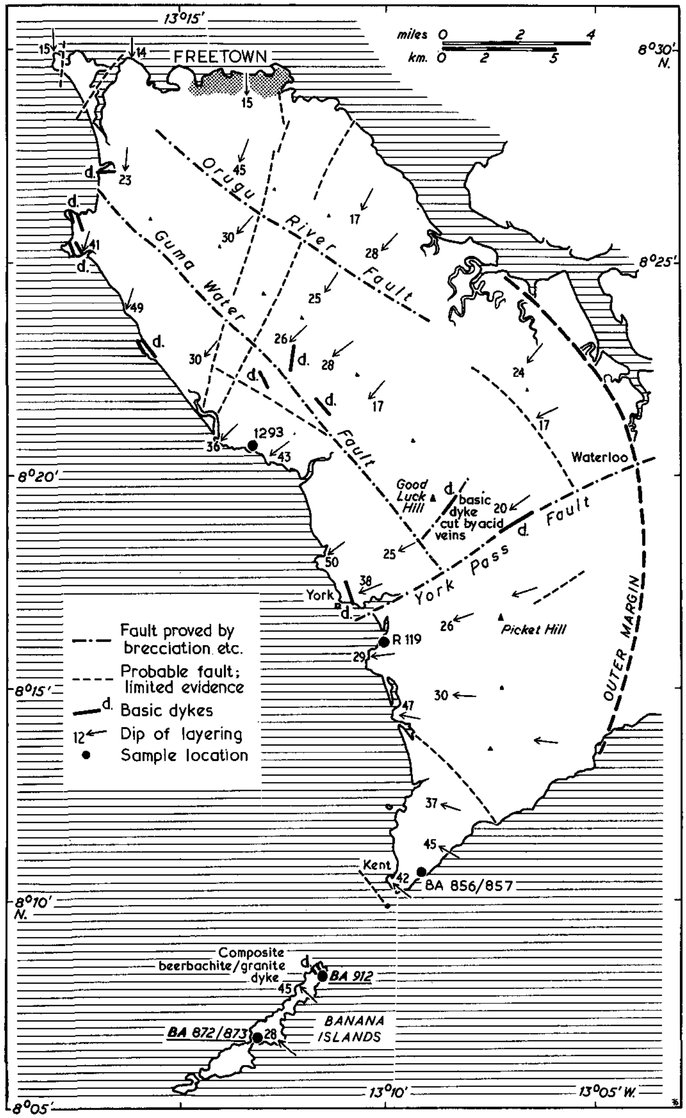

FIG. 1. Outline map of the Freetown Complex, Sierra Leone, showing locations of the analysed samples of vein material.

reliable and that the spectrum of younger dates might reflect ^{40}Ar loss. Andrews-Jones (1968), however, argued in favour of a Mesozoic age for the Freetown Complex because an amphibole separated from the country rock just below the basal contact gave a K-Ar age of 306 ± 20 Ma. Andrews-Jones concluded that the youngest ages published by Snelling (1966) were likely to be the most realistic and that the older ages were probably due to the presence of excess ^{40}Ar. K-Ar ages of 189 ± 10 Ma and 235 ± 10 Ma determined at I.G.S. for whole rock samples of two members of the WNW dyke swarm that cut the Freetown Complex led Andrews-Jones to suggest that the age of the major intrusion probably lies between 235 Ma and 306 Ma. Briden, Henthorn, and Rex (1971) determined K-Ar ages for three rock samples from the Freetown Complex in the range 169 ± 10 Ma to 198 ± 8 Ma and suggested that the mean of their results, about 184 Ma, could be taken as a minimum age for the major intrusion. These authors also determined palaeomagnetic pole positions for the Freetown Complex and showed that these are only matched in Africa by poles of Triassic to Lower Jurassic age (i.e. about 225 Ma–165 Ma). It is clear that although there is now little doubt that the Freetown Complex is of Mesozoic age, the intrusion has not been dated directly with any degree of accuracy. The conventional K–Ar ages from 190 Ma to 235 Ma for dykes cutting the intrusion provide only a crude younger age limit, while palaeomagnetic studies (Briden *et al.*, 1971) on the major intrusion are consistent with an age somewhere in the approximate range 165 Ma–225 Ma. The purpose of this paper is to present a new approach to dating the Freetown Complex (Rb–Sr whole-rock dating of acid veins), which we believe yields a definitive result.

Rb–Sr studies of acid veins in the Freetown Complex. Because of the difficulties, outlined above, of interpreting K–Ar age studies in this case, and the analytical problems of determining precise Rb–Sr ages on young basic rocks, it was decided to attempt a Rb–Sr study of the acid veins associated with the Freetown Complex (Wells, 1962, p. 95). These acid veins are sparsely distributed throughout the intrusion and are typically about 1 cm in thickness although rarely they reach a thickness of about 10 cm. They are generally parallel-sided and are seen to extend for distances of tens of metres along planes dictated by the joint directions of the major intrusion. Beerbachite dykes (Wells, 1962, p. 91) are found, especially in the Banana Islands (fig. 1) and are probably related to the magma forming the normal basic ('doleritic') dykes of the Complex. The difference between these two types of basic intrusion is believed to be due to the condition of the host rock at the time of injection. Whereas the dolerite dykes have chilled margins and were thus intruded when the local host rocks were consolidated and relatively cool, the beerbachites appear to have been injected into hot rock, probably under conditions equivalent to those of pyroxene hornfels metamorphism, at temperatures in excess of 800 °C (Turner, 1968). The composite nature of the beerbachite dyke, granitic vein, and apophyses shown in fig. 2 demonstrates conclusively that formation of the acid and basic materials was essentially contemporaneous. Thus it appears that the acid veins—which may occur in association with beerbachite or dolerite (fig. 1 south of Good Luck Hill), or more commonly in gabbroic and troctolitic rocks—were formed within relatively early stages of the cooling history of the Complex. Petrological studies, described later, show that the acid veins are formed principally *in situ* by reactions between hydrous fluids and the host rock.

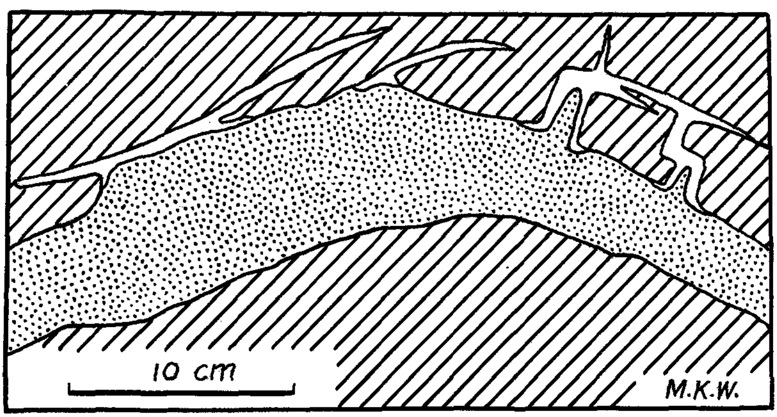

FIG. 2. Detail of composite vein in the Banana Islands showing acidic apophyses marginal to contemporaneous granulitized beerbachite, Ba 912. Gabbro shown ruled; beerbachite stippled; acidic apophyses blank.

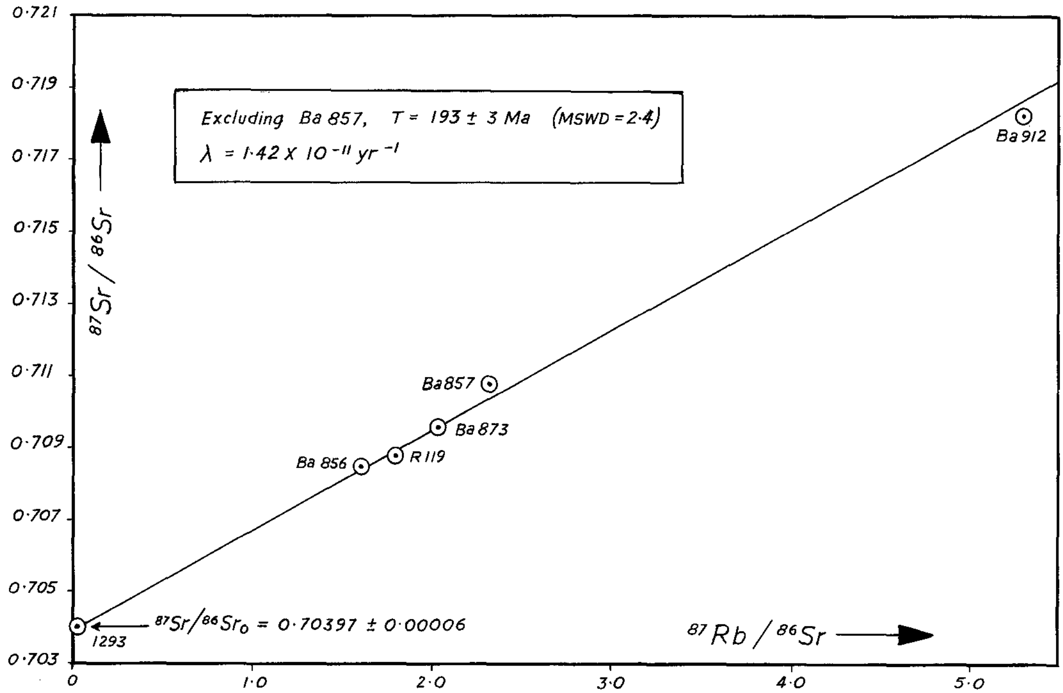

FIG. 3. Rb–Sr whole rock isochron of the analysed material.

Occasionally (e.g. specimen R 119) field relations indicate localized intrusion of the acid vein material.

Five samples of the acid veins and a sample of a basic pegmatite from the localities indicated in fig. 1 have been analysed for Rb and Sr contents and Rb/Sr and $^{87}Sr/^{86}Sr$ ratios by standard XRF and mass spectrometric methods (Pankhurst and O'Nions, 1973), using fine powder prepared from about 60 grammes of vein material. The analytical results are shown in Table I and an isochron diagram (fig. 3). A least-squares regression (Williamson, 1968) of the acid rocks alone yields an age of 194 ± 23 Ma and an initial $^{87}Sr/^{86}Sr$ ratio of $0{\cdot}70405 \pm 0{\cdot}00068$. The quoted confidence limits (2-sigma) include scatter in excess of analytical error.

TABLE I. *The results of Rb, Sr, and $^{87}Sr/^{86}Sr$ analyses*

Sample no.	Rb	Sr	$^{87}Rb/^{86}Sr$	$^{87}Sr/^{86}Sr$†
Acid rocks	ppm(±10%)		±1%	
R 119	79	126	1·800	0·70883 ± 0·00003
Ba 912	272	148	5·306	0·71829 ± 0·00005
Ba 856	126	226	1·615	0·70848 ± 0·00002
Ba 873	142	201	2·043	0·70960 ± 0·00003
Ba 857	113	141	2·321	0·71080 ± 0·00003
Basic pegmatite				
1293	3	364	0·029	0·70403 ± 0·00003

† ± standard error, normalized to $^{86}Sr/^{88}Sr = 0{\cdot}11940$.

Clearly the initial $^{87}Sr/^{86}Sr$ ratio of the basic pegmatite (Table I) cannot be distinguished from that of the acid-vein material and we feel justified in including this sample in a second regression. This yields an age of 197 ± 9 Ma and an initial ratio of $0{\cdot}70396 \pm 0{\cdot}00019$. Most of the remaining scatter can be attributed to sample Ba 857, which in thin section appears extensively weathered in contrast to the other specimens. This data point is excluded in a third regression and the remaining points fit a straight line within experimental errors giving a MSWD of 2·4 and an age and initial ratio of 193 ± 3 Ma and $0{\cdot}70397 \pm 0{\cdot}00006$. We conclude that this regression provides the most reliable estimate of both age and initial $^{87}Sr/^{86}Sr$ ratio. The mean value obtained for the Eimer and Amend standard during this work was $0{\cdot}70808 \pm 0{\cdot}00001$ and thus the initial $^{87}Sr/^{86}Sr$ ratio for both basic and acid components of the Freetown Complex should be quoted as $0{\cdot}70389 \pm 0{\cdot}00006$ relative to an Eimer and Amend value of 0·70800.

Discussion and conclusions

Rb-Sr whole rock dating of acid veins. When combined with the field evidence cited above, that the acid veins are contemporaneous with the early stages of cooling of the major intrusion, this appears to be a straightforward method of dating basic intrusions that are difficult to date precisely using any other approach. The initial $^{87}Sr/^{86}Sr$ ratio of the acid-vein material in the Freetown Complex is constant over the wide area from which the samples were taken (fig. 1). Possible reasons for this and petrographic criteria that would be useful in identifying similar veins in other basic intrusions are discussed below.

Origin of the acid veins in the Freetown Complex. The veins consist principally of quartz and feldspar, which frequently display graphic intergrowths. However, the proportions of quartz and feldspar are variable as can be appreciated from the norms calculated from the major element analyses presented in Table II.

Although the margins of the veins appear sharp in hand specimen, they are less obvious in thin section. The margins are often marked by a change in the composition of the feldspars and by breakdown of the ferromagnesian minerals along an approximately straight boundary (figs. 4 and 5). Pericline twinning has been induced mechanically in the plagioclase where it projects from the gabbro into the vein and the feldspar changes from coarsely lamellar plagioclase in the gabbro to altered plagioclase with a fine lamellar structure plus orthoclase in the vein (figs. 4 and 5). Within the veins it is possible to find plagioclase with a wide rim altered to orthoclase, and orthoclase grains showing relict plagioclase twin lamellae (figs. 4 and 5). Occasionally (e.g. sample Ba 912) stringers of quartz run parallel to the vein margin and are particularly abundant at the centre of the vein.

Electron-microprobe analyses of nine minerals occurring near the vein boundary in sample Ba 856 are given in Table III. The locations of the spots analysed by the electron microprobe are indicated on the microphotograph of fig. 4. These analyses clearly show that some of the vein feldspars have been formed by alteration of labradorite from the host rock. The clear labradorite of the host rock shows a balanced molecular formula (Table III, analysis 1). The same is true of the dominant clear orthoclase (analysis 2) and the orthoclase component of the perthitic feldspar (analysis 6). By contrast, the plagioclase component of the perthite contains a significant amount of excess silica (analysis 7). The original feldspars of host composition, represented by analysis 1, have been progressively altered towards a silicified albite composition (analyses 3, 4, and 8) principally by the addition of silica and loss of calcium. Analysis 8 at the rim of the altered plagioclase corresponds to one of the two components of a symplectic intergrowth, which has a myrmekitic appearance and occurs in a zone between the plagioclase and the perthitic feldspar (analyses 6 and 7). The other component (analysis 9), lying closer to the perthite, has the composition of a mixture of 49 % quartz and 51 % andesine.

TABLE II. *Major-element analyses of the vein material*

	R 119[1]	Ba 912[1]	Ba 873[1]	Ba 857[1]	Ba 856[1]	1293[2]	316[3]
SiO_2	70·07	60·52	59·76	64·36	61·78	46·38	77·38
Al_2O_3	14·31	12·86	16·58	14·03	16·33	17·84	11·83
Fe_2O_3	2·40	2·15	2·76	3·03	2·40	1·77	0·76
FeO	1·25	5·60	3·33	2·73	3·33	5·63	0·30
MgO	0·24	4·97	3·37	3·64	3·66	5·91	0·08
CaO	1·47	3·67	6·38	4·94	6·76	16·05	0·60
Na_2O	3·81	2·91	2·90	2·07	2·46	2·63	2·40
K_2O	4·75	4·67	3·22	3·73	2·81	0·17	6·00
H_2O+	0·27	0·95	0·76	0·97	0·65	0·46	0·35
TiO_2	0·34	0·40	0·28	0·36	0·29	1·57	0·18
P_2O_5	0·07	0·12	0·00	0·00	0·00	0·16	0·01
MnO	0·04	0·11	0·07	0·06	0·07	0·13	Tr.
CO_2	0·91	1·41	0·64	0·33	0·43	0·40	0·00
Total	99·93	100·34	100·05	100·25	100·97	99·10	100·04
CIPW norms							
qz	28·92	11·05	13·10	22·99	17·46	0·00	39·00
or	28·06	27·59	19·02	22·04	16·60	1·00	35·45
ab	32·22	24·61	24·53	17·51	20·81	13·51	20·29
an	1·08	8·24	22·72	17·98	25·22	36·37	2·99
di	0·00	0·22	3·93	3·53	4·50	32·00	0·00
hy	0·60	20·32	10·00	9·41	10·73	0·00	0·20
mt	3·17	3·12	4·00	4·39	3·48	2·57	0·43
hm	0·21	0·00	0·00	0·00	0·00	0·00	0·47
il	0·65	0·76	0·53	0·68	0·55	2·98	0·33
ap	0·17	0·29	0·00	0·00	0·00	0·38	0·02
cc	2·07	3·21	1·46	0·75	0·98	0·91	0·00
cor	2·51	0·00	0·00	0·00	0·00	0·00	0·29

1. Acid veins. 2. Basic pyroxene pegmatite. 1 and 2. Atomic absorption analysis by H. M. Blauer on powders split from the samples used for Rb–Sr dating. 3. Wet-chemical analysis by Harwood quoted by Junner (1930) on vein material from south of Good Luck Hill (fig. 1), Tr = Trace. Includes ZrO_2 = 0·01, BaO = 0·04, V_2O_5 = 0·01, S = 0·01, H_2O = 0·08, and trace Cl and Li_2O.

Iron–titanium oxide grains in the veins show textural similarity with those found in the gabbro (Wells, 1962; Bowles, 1976, 1977). Electron-microprobe analysis reveals the presence of silicon and some calcium along the planes of alteration between formerly exsolved ilmenite lamellae and host magnetite, although both are now extensively altered to maghemite. Interstitial patches of carbonate are frequently found in the veins. The mineralogy of the apophyses of the veins is directly controlled by the phases in the gabbro through which the apophyses pass. The apophyses are composed of quartz where they pass through host olivine or pyroxene, but of orthoclase (occasionally containing relict twin lamellae in approximate optical continuity with those in the host plagioclase) where they cut plagioclase (fig. 5). In the gabbro, close to the veins, plagioclase crystals show myrmekitic textures, and symplectic textures occur between olivine and magnetite. These textures are not found in the host rock away from the veins.

Thus the textural evidence strongly suggests that the veins have formed essentially *in situ* and that alteration of minerals in the host gabbro has provided much of the vein material. In general, there is no evidence for bulk transport of material along the veins, as shown by many relict and undisplaced features such as twin lamellae described above. Nevertheless, during the formation of the fractures, which are now infilled with acid material, some crystals were detached from the host rock, becoming incorporated into the vein and extensively altered (fig. 4), and some

TABLE III. *Electron-microprobe analyses of the felsic minerals of specimen Ba 856*

	Host	Vein							
	1	2	3	4	5	6	7	8	9
SiO_2	52·53	64·13	60·45	60·10	99·69	63·44	62·44	62·81	82·55
TiO_2	0·07	0·03	0·02	0·03	0·02	0·03	0·02	0·03	0·03
Al_2O_3	30·12	19·34	24·78	24·51	0·03	19·22	24·09	24·20	9·42
V_2O_3	0·01	0·01	0·00	0·00	0·00	0·02	0·00	0·00	0·00
FeO*	0·26	0·10	0·26	0·22	0·01	0·11	0·12	0·19	0·31
MnO	0·01	0·00	0·00	0·02	0·00	0·01	0·00	0·01	0·00
MgO	0·04	0·02	0·03	0·06	0·00	0·04	0·05	0·00	0·22
CaO	12·67	0·34	6·10	5·72	0·00	0·07	4·57	5·28	2·15
NiO	0·00	0·00	0·00	0·00	0·00	0·02	0·01	0·00	0·00
Na_2O	3·93	2·85	6·29	4·43	0·00	0·99	6·34	4·98	4·63
K_2O	0·21	11·62	0·39	0·20	0·01	15·07	0·65	0·15	0·08
Total	99·85	98·44	98·32	95·29	99·76	99·02	98·29	97·65	99·39
			Molecular formulae $(K, Na, Ca)_Z$ $(Al, Si)_Y$ O_{32}						
Si	9·538	11·851	10·875	11·026	—	11·820	11·154	11·216	13·960
Ti	0·010	0·004	0·003	0·004	—	0·004	0·003	0·004	0·004
Al	6·446	4·212	5·254	5·300	—	4·221	5·072	5·093	1·877
Total Y	15·99	16·07	16·13	16·33	—	16·05	16·23	16·31	15·84
V	0·001	0·001	—	—	—	0·003	—	—	—
Fe	0·039	0·015	0·039	0·034	—	0·017	0·018	0·028	0·044
Mn	0·002	—	—	0·003	—	0·002	—	0·002	—
Mg	0·011	0·006	0·008	0·016	—	0·011	0·013	—	0·055
Ca	2·465	0·067	1·176	1·124	—	0·014	0·875	1·010	0·390
Ni	0·001	—	—	—	—	0·003	0·001	—	—
Na	1·384	1·021	2·194	1·576	—	0·358	2·196	1·724	1·518
K	0·049	2·739	0·089	0·047	—	3·582	0·148	0·034	0·017
Total Z	3·95	3·83	3·51	2·80	—	3·99	3·25	2·80	2·02
Quartz %	1·1	4·2	13·0	31·4	100·0	0·6	19·9	31·4	49·0
Feldspar %	98·9	95·8	87·0	68·6	0·0	99·4	80·1	68·6	51·0
Feldspar compositions									
Ab%	35·1	26·5	62·6	56·4	—	9·0	67·5	61·7	75·0
An%	63·7	2·3	34·9	42·0	—	1·0	27·9	37·1	24·2
Or%	1·2	71·2	2·5	1·6	—	90·0	4·6	1·2	0·8

1. Host rock plagioclase. 2. Clear orthoclase. 3. Altered plagioclase. 4. Rim to altered plagioclase. 5. Quartz. 6. Orthoclase product of alteration. 7. Relict altered plagioclase. 8. Altered plagioclase rim, partner to 9. 9. Myrmekitic siliceous rim to plagioclase.

* Total Fe, given as FeO.

The analytical methods are described in Bowles (1975) and the quantitative microprobe correction program of Mason, Frost, and Reed (1969) was used. The locations of these numbered analyses are indicated in fig. 4 and Cr_2O_3 = 0·00 wt % throughout. The molecular formulae are calculated to 32 oxygen atoms and show considerable excess silica in the altered feldspars where 'Total Y' is greater than 16 and 'Total Z' is much less than 4. Since the alteration is so distinct the molecular formulae have been split into quartz+feldspar by removing silica from the formulae until an exact 16:4 balance is achieved. The resulting quartz and feldspar proportions are shown above, together with the Ab-An-Or proportions of the remaining feldspar.

localized transfer of material did occur. It appears that as joint planes formed during the early cooling of the Freetown Complex, an acid fraction was able to differentiate from the surrounding solid gabbro to occupy the progressively opening joints. The fractures in the gabbro were probably filled by a hydrous fluid phase, which was responsible for the oxidation of the iron–titanium oxides, the serpentinization of the host olivine, and the formation of the

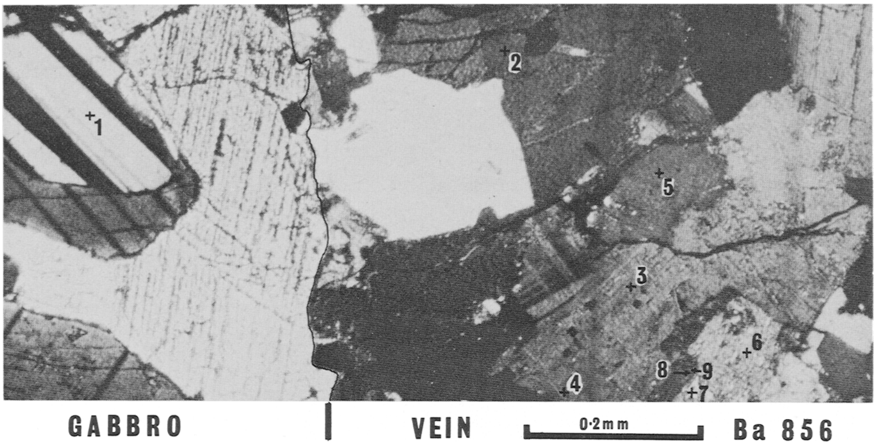

FIG. 4. Annotated microphotograph of Ba 856 showing the location of the points analysed by electron-microprobe.

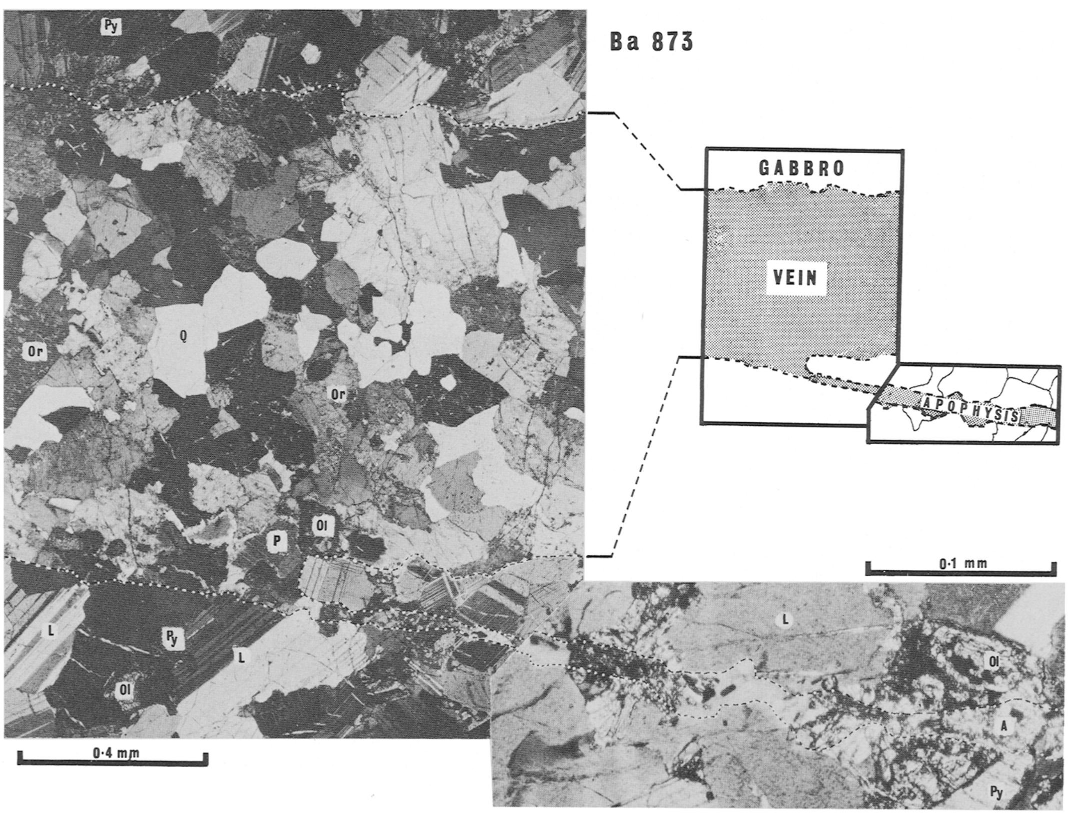

FIG. 5. A typical acid vein cutting gabbro. Fresh labradorite (L) in the gabbro contrasts with relict plagioclase (P) and olivine (Ol) at the edge of the vein. Olivine (Ol) and pyroxene (Py) occur in the gabbro whilst quartz (Q) and orthoclase (Or) are seen principally in the centre of the vein. A narrow apophysis leads away from the vein and shows features identical with those shown by other incipient veins, which parallel the principal vein shown here. The continuation of the apophysis is shown at a larger scale to illustrate the close control that the mineralogy of the host material exerts on the mineralogy of the vein. Where the apophysis passes through labradorite (L) a felsic phase is formed whilst a mafic phase (? amphibole, A) is found where the apophysis cuts olivine and pyroxene.

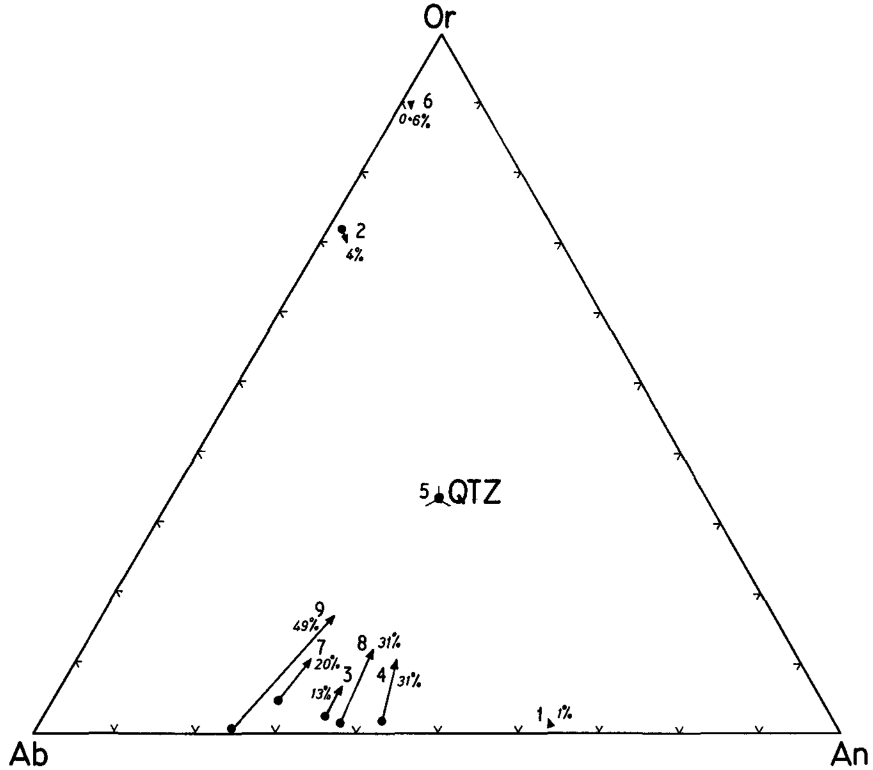

FIG. 6. Quartz–albite–orthoclase–anorthite tetrahedron projected on to its feldspar base. The arrowheads represent the projected composition of the feldspar with excess silica included, whilst the 'tail' of the arrow represents the composition of the feldspar component alone plotted on the feldspar base. The arrows are, in effect, tie lines whose length corresponds to the excess silica content (see Table III), the percentage of which in the total mixture is shown.

interstitial carbonates. The process by which the veins became occupied by felsic minerals is, in many ways, similar to the diffusion processes that fill tectonic veins associated with pressure-solution fabrics (Elliot, 1972; Kerrich *et al.*, 1976). The transfer of components from the host rock into low-pressure regions associated with incipient fractures occurs by diffusion of ions and molecular species through a relatively static pore fluid or along grain boundaries. The chemical potential that drives the diffusion is a response to stress gradients (Kerrich, Beckinsale, and Shackleton, private communication), which existed locally where the hydrostatic (P_{fluid}) pressure was less than P_{rock}. This mechanism is consistent with the fact that the initial $^{87}Sr/^{86}Sr$ ratios for both acid and basic rocks have an identical low value ($0{\cdot}70389 \pm 0{\cdot}00006$) suggesting a genetic relationship. There is no indication that formation of the veins has involved contamination by more radiogenic strontium such as might have been derived from the local Precambrian basement.

Excess ^{40} Ar in the Freetown Complex and Liberian dykes. There is no doubt that the older K–Ar ages reported by Snelling (1966) for the basic pluton are discrepantly old, presumably because of the presence of substantial levels of excess ^{40}Ar in the plagioclase and pyroxene separates analysed. On the other hand, the age of intrusion determined here agrees closely with the youngest of the conventional K–Ar ages. Dalrymple, Grommé, and White (1975) have described detailed K–Ar studies of the coast-parallel dyke swarm in Liberia. Whole rock and plagioclase samples of dykes intruding Precambrian crystalline rocks yielded a range of conventional K-Ar ages from 190 Ma to 1221 Ma. However, whole rock and plagioclase samples of dykes and sills intruding Palaeozoic sediments (the Paynesville Sandstone) yielded a restricted range of conventional K-Ar ages from 177 Ma to 196 Ma. Furthermore, samples of the dykes intruding Precambrian rocks that yielded old K-Ar ages gave 'saddle-shaped' $^{40}Ar/^{39}Ar$ release patterns with the nearest approach to a plateau represented by the intermediate temperature minimum, corresponding to less than 300 Ma. One whole rock sample and one separated plagioclase from an intrusion in the Paynesville Sandstone, which yielded a conventional K-Ar age of 182 ± 4 Ma, gave $^{40}Ar/^{39}Ar$ plateau ages of $189{\cdot}0 \pm 4{\cdot}4$ Ma and $191{\cdot}0 \pm 3{\cdot}4$ Ma respectively. Dalrymple, Grommé, and White (1975) concluded, therefore, that the dykes intruding the Precambrian crystalline rocks contain large and variable amounts of excess ^{40}Ar, and that the whole coast-parallel dyke swarm was intruded at about 190 Ma ago. This is in close agreement with the conclusion presented here for the age of the Freetown Complex. Excess ^{40}Ar in both the Freetown Complex and the dyke swarm is most likely to have been derived from degassing of the local Precambrian crystalline country rock.

Implications for Mesozoic palaeomagnetic stratigraphy. It appears from the work of Briden, Henthorn, and Rex (1971) that at least one polarity reversal occurred during the cooling of the Freetown Complex. It was found that the stable thermoremanent magnetism has reversed polarity at localities stratigraphically within about 1 km of the base of the intrusion, but normal polarity at localities elsewhere in the intrusion (except for one sample with a low NRM/TRM ratio, which they were inclined to relate to a second polarity transition). These authors suggested that the definite polarity reversal reflected *upward* migration of the Curie isotherm and thus the transition was from reversed to normal polarity. However, at Curie temperatures characteristic of the magnetites in gabbros (about 570 °C) it would seem more likely that the intrusion would have cooled from the upper surface downwards. The polarity reversal recorded in the Freetown Complex is, therefore, probably from normal to reversed. It is interesting to note that a comparable transition, marking the beginning of the Nuanetsi Zone of reversed polarity, is recorded in the Marangudzi ring complex of Rhodesia (McElhinny and Burek, 1971). Foland and Henderson (1975) have recently determined a Rb-Sr whole rock isochron age of 186 ± 3 Ma for this complex ($\lambda = 1{\cdot}42 \times 10^{-11} yr^{-1}$). The ages of the Freetown Complex and the Marangudzi Complex differ by 1 Ma at the limits of error. However, it is possible that systematic errors of this magnitude (0·5 %) could be present in both age determinations and at this stage it appears reasonable to conclude that these two intrusions were essentially synchronous and that the mean age, about 190 Ma, is the best available estimate for the beginning of the Nuanetsi Zone. Most of the Liberian dykes have normal polarity, and a few are reversed (Dalrymple *et al.*, 1975). Furthermore, the palaeomagnetic pole derived from these dykes is in close agreement with that derived from other Mesozoic igneous rocks from Africa, including the Freetown Complex (Dalrymple *et al.*, 1975). Thus both geochronological and palaeomagnetic studies strongly suggest contemporaneous intrusion of the Freetown Complex and the dyke swarm. The coast-parallel structure of the dyke swarm suggests a link between this magmatic event and the regional stress pattern that led to the separation of the continents in this region.

Acknowledgements. The $^{87}Sr/^{86}Sr$ ratios were measured at the Department of Geology and Mineralogy, University of Oxford, while R. J. P. was in receipt of a N.E.R.C. Research Assistantship. Isotope studies at Oxford are supported by N.E.R.C. through a research grant to Drs. Moorbath and N. H. Gale to whom we are also indebted. This paper is published with permission of the Director, Institute of Geological Sciences.

REFERENCES

Andrews-Jones (D. A.), 1968. *Petrogenesis and Geochemistry of rocks of the Kenema district, Sierra Leone.* Unpublished Ph.D. thesis, University of Leeds.

Bowles (J. F. W.), 1975. *Rep. Inst. Geol. Sci.* **No. 75/9** [M.A. 76-82].

—— 1976. *Mineral. Mag.* **40,** 703-14 [M.A. 76-3633].

—— 1977. Ibid. **41,** 103-9.

Briden (J. C.), Henthorn (D. I.), and Rex (D. C.), 1971. *Earth Planet. Sci. Lett.* **12,** 385-91.

Dalrymple (G. B.), Grommé (C. S.), and White (R. W.), 1975. *Bull. Geol. Soc. Am.* **86,** 399-411.

Elliot (D.), 1972. Ibid. **84,** 2645-64.

Foland (K. A.) and Henderson (C. M. B.), 1975. *Earth Planet. Sci. Lett.* **29,** 291-301.

Junner (N. R.) and Harwood (H. F.), 1930. *Rep. 15th Sess. Internat. geol. Cong., S. Afr.*, 1929, No. **43,** 417-33.

Kerrich (R.), Beckinsale (R. D.), and Durham (J. J.), 1977. *Tectonophysics,* **38,** 241-57.

———— and Shackleton (N. J.). Unpublished MS.

McElhinny (M. W.) and Burek (P. J.), 1971. *Nature,* **232,** 98-102.

Mason (P. K.), Frost (M. T.), and Reed (S. J. B.), 1969. *Natl. Phys. Lab. (I.M.S.)*, Rep. 2.

Pankhurst (R. J.) and O'Nions (R. K.), 1973. *Chem. Geol.* **12,** 127-36.

Snelling (N. J.), 1966. *Inst. Geol. Sci. Ann. Rep. for 1965.* **II,** 44-57.

Turner (F. J.), 1968. *Metamorphic Petrology,* McGraw-Hill, New York.

Wells (M. K.), 1962. *Geol. Surv. Dept. Sierra Leone, Short Paper,* No. **9.** (*Bull. Overseas Geol. Min. Res., Suppl.* No. **4**) [M.A. **16**-305].

Williamson (J. H.), 1968. *Canad. Journ. Phys.* **46,** 1845-7.

[*Revised manuscript received 20 April 1977*]