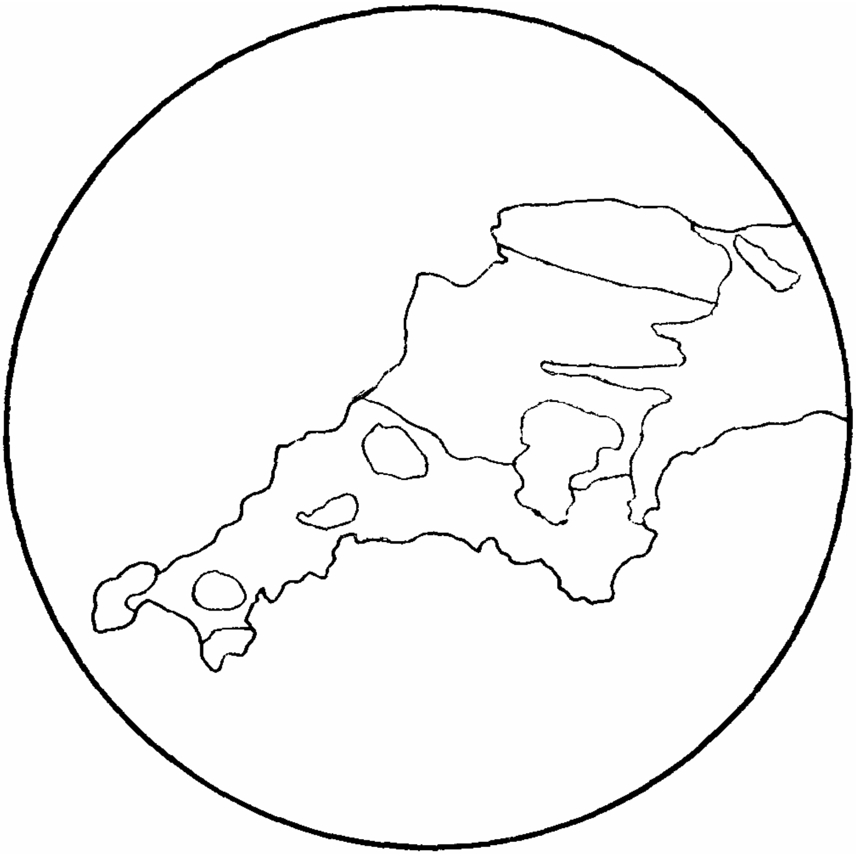
PROCEEDINGS

OF THE

USSHER SOCIETY

VOLUME FOUR

PART ONE



1977

THE USSHER SOCIETY

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Edited by
A. WHITTAKER

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Conference of the Ussher Society held at Torquay, January 1977

Chairman's Report

Having decided to meet again at the Queen's Hotel, Torquay, for the Annual Conference, members and guests arriving for field trips on Sunday, 2nd January, 1977, must have appreciated the crisp, clear and sunny day provided for them. This in marked contrast to the weather at early meetings. One party, led by Dr. C.T. Scrutton, visited a number of localities in the Devonian limestones and other facies in the Torquay promontory. Another party, led by Dr. J.C. Laming, visited exposures of the New Red Sandstone, especially those revealed by new motorway development. Members, thus refreshed by an excellent *aperitif,* were put in a proper mood for the conference itself which commenced the next day.

In the programme for 3,4 January over forty contributors were scheduled in a programme of twenty-five talks. Pride of place as guest speaker was given to Dr. C. Halls who initiated the programme with his lecture on 'Tin deposits - some comparative aspects'. This was followed, as usual, by talks over a very wide range of geological interests in as varied a programme as the Society has ever had. Dr. R.T. Taylor acted as Conference Secretary, and much of the success was due to him; again the University of Exeter helped with the provision of equipment.

General problems encountered by the Society at present reflect the inflationary times. There is a need for constant review of subscription and of publication costs. The back stocks of any Society's publications in such circumstances represent a significant capital reserve. Unfortunately, too few copies of some parts of volume three were printed so that they are about to run out. Since this would mean that the opportunity to sell complete back runs would be lost, a modest programme of reprinting has been agreed. This must precede any large-scale attempt at a widespread international sale and subscription campaign for which, otherwise, the time is ripe. In future, negatives of parts will be kept so that

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reprinting will be cheaper and easier and there will be no need to tie up too much capital in back stock.

It is a pleasure to record my indebtedness to the Society Officers for their unstinted and enthusiastic work and to their, and my, predecessors who have established so efficient a pattern. Our Society is exceptional among regional societies in the vigour and range of the activity of its members. Fortunately there remain a sufficient number of geological problems in south-west England to keep us active for a long time to come.

The Society would like to thank Richard Scrivener for compiling the index to Volume three of the *Proceedings*.

Michael House
July 1977

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Mineralogy and paragenesis at the Mount Wellington Mine (Abstract):

by Y.A. Kettaneh and J.P.N. Badham

The Mount Wellington Mine lies some 7 km west of Truro, 3 and 4 km respectively from the nearest outcrops of the Carnmarth and Carnmenellis granites. The mine, which is on the westerly extension of the Wheal Jane lodes, is within lightly metamorphosed Lower Devonian phyllites. Workings in the area were initiated in the eighteenth century but the present phase of mining was initiated in 1964 when, of sixty holes drilled, forty-six intersected the important Number 1 lode beneath a gently dipping elvan sheet. In 1968 the property was placed under the management of Cornwall Tin and Mining Limited and was brought into production in 1975.

The main production comes from the Number 1 lode but two other lodes (2 and 3), with the same strike, dip more steeply and join the Number I lode in the elvan footwall. These lodes are generally smaller but of higher grade. All the lodes have a similar mineralogy and paragenesis. In each the two early phases of mineralisation replace brecciated fault zones and the two later phases fill dilatant fractures in the earlier material. All four phases are thought to have occurred in relatively rapid sequence and above a temperature of 300°C. These lodes are cut by variously mineralised caunter lodes and cross-courses which are probably considerably younger. A smaller lode - the Hot Lode - contains high copper grades and may correlate with the latter phases of primary mineralisation.

The lodes contain Fe, Zn, Sn, Cu, As, Ti and lesser Pb, Ag, Au, W, Sb and Bi. Tin, copper and zinc concentrates are profitably produced. The mineralogy and paragenesis are shown on the accompanying figure. A full description will be published elsewhere. The elvan and the main mineralisation phases were emplaced in two sets of faults which developed around the cooling Carnmenellis granite cupola probably in response to fluid overpressure. The efficiency of the faults as deposition sites is suggested by the coincidence of elvan and lodes and by the spatially telescoped paragenesis.

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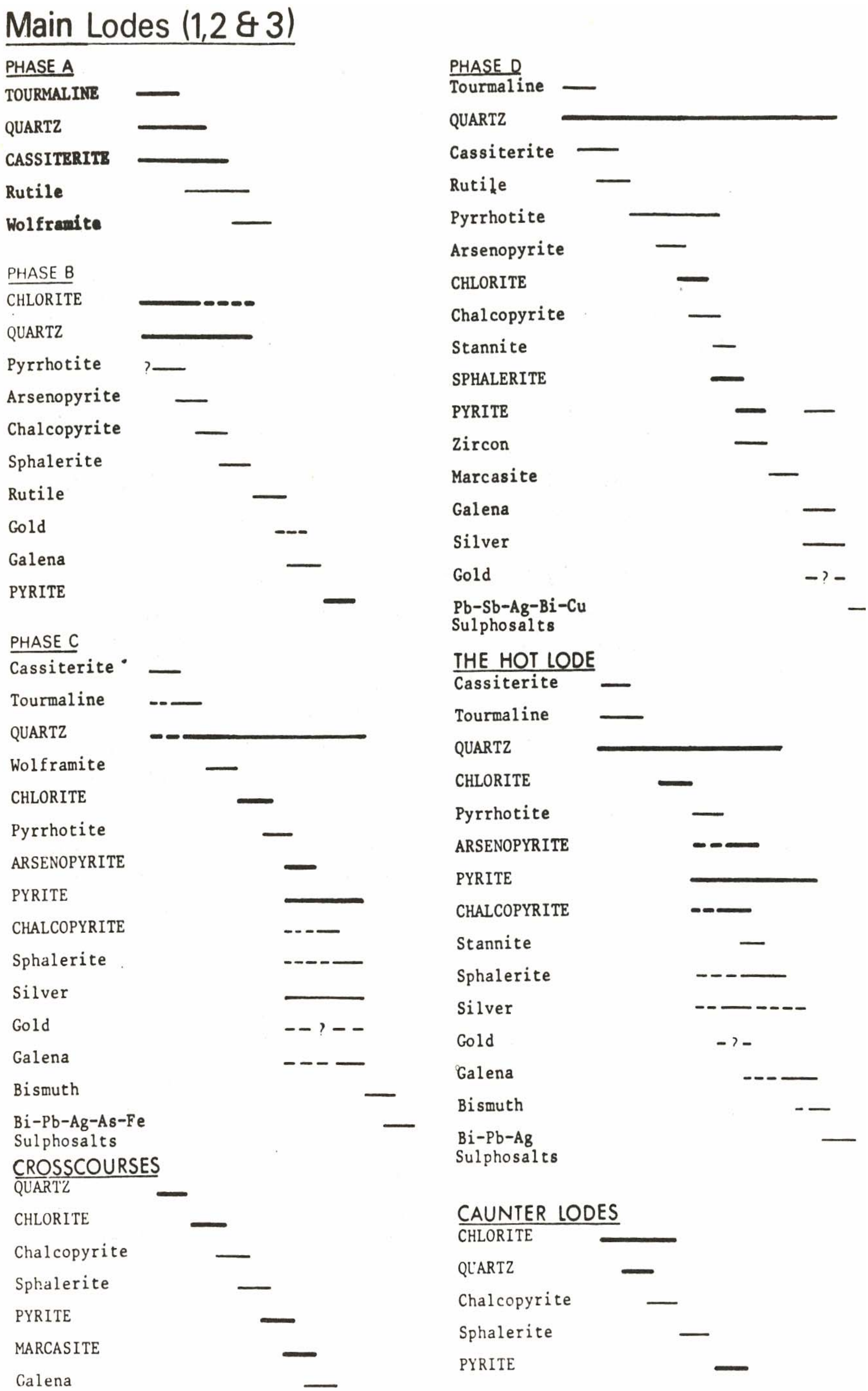

FIGURE 1 Mineralogy and paragenesis at Mount Wellington. Major minerals in each phase of mineralization are indicated in upper case

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SOME NOTES ON THE MINERALISATION OF THE DARTMOOR GRANITE

by R.C. Scrivener, B.V. Cooper and O.A. Baker

Abstract. Examination of a large number of ore specimens forms the basis of a mineralogical review of the Dartmoor metalliferous veins. Paragenesis is discussed and the specular hematite mineralisation is considered to be a late-stage phenomenon.

1. Introduction

Information about the distribution, mineralogy and paragenesis of the metalliferous hydrothermal veins of the Dartmoor Granite is rather limited. In the past, attention has been focused on the more productive districts, namely, the tin mining complex of the Birch Tor area (MacAlister, 1909, Dines, 1956, p. 720 and Broughton, 1968) and the Hennock--Lustleigh micaceous hematite mines (Dines, 1956, p. 725, Henson, 1956). A short general account dealing with mineralogy and lode trend is given by Brammall (1928).

In this study, specimens of ore and wallrock collected from a wide range of localities have formed the basis of an attempt to assess the general nature of the hydrothermal veins and, in particular, their paragenesis and distribution. Unfortunately, *in situ* exposures of the lodes are rare and it is necessary to rely on material from mine dumps in many cases.

2. Mineralogy

Cassiterite is the most widely distributed economic mineral in the Dartmoor lodes. Varying in colour from dark reddish brown to pale brownish yellow, it commonly exhibits striking zoning when seen in thin section. It is normally well crystallised and of subhedral habit. Cassiterite is invariably associated with quartz and acicular blue tourmaline which, together with brown tourmaline, feldspar and hematite, form the major gangue components.

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The specular variety of hematite is of widespread occurrence and is a major component of the lodes of Central Dartmoor (MacAlister, 1909). It often exhibits interesting replacement textures, invading kaolinised wallrock at Kelly Mine (SX 794 817) and forming pseudomorphs after pyrite at a site near Golden Dagger Mine (SX 679 803). The fine-grained, micaceous variety of hematite, formerly used in the manufacture of paint, is confined to north-eastern Dartmoor, where it forms the major constituent of narrow, irregular veins, close to the granite contact. Associated minerals are tourmaline, quartz, kaolin and pyrite. A specimen of washed micaceous hematite from Great Rock Mine (SX 827 815), subjected to XRD examination (B.J. Sheldon, pers. comm.) shows that the sample is nearly pure ferric oxide, with a trace of muscovite as the only impurity. Earthy red hematite and minor quantities of kidney ore are very widely distributed. Magnetite is present in vein material from a site near Stoneslade Tor (SX 712 783).

Sulphides are rare in the Dartmoor lodes and, apart from traces of pyrite and chalcopyrite in the central district, are mainly confined to ores from sites close to the contact. Well-formed pyrite crystals up to 40 mm across are present at Great Rock Mine. Arsenopyrite and sphalerite are present in minor quantities at Crownley Parks Mine (SX 760 764). An unusual association of chalcocite with quartz and purple fluorite occurs at Wheal Cumpston (SX 672 723).

Chalcedonic epimorphs of fluorite have been noted at Ringleshutes Mine (SX 677 698), and quartz pseudomorphs after the same mineral at Huntingdon Warren (SX 671 670). Wolframite, although an important mineral at Hemerdon Mine (SX 573 584), has only been noted in minor amounts elsewhere, particularly at Eylesbarrow Mine (SX 599 682). Traces of gold have been noted at various localities (Brammall, 1926).

3. Paragenesis

The paragenesis of tin ores from the central and north-eastern parts of Dartmoor is exemplified by consideration of the following genetic history of a specimen from Great Weeke Mine (SX 715 876).

(i) Initiation of fracture and deposition of quartz and aggregates of dark-brown tourmaline.

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(ii) Second Stage of fracturing, producing a breccia which is cemented by deposition of blue acicular tourmaline, cassiterite and quartz. The cassiterite is bracketed by the growth of tourmaline.

(iii) Specular hematite is deposited along further fractures, in places replacing quartz.

(iv) Late quartz is deposited.

In the general case, sulphides and fluorite are deposited after stage (ii) but before the introduction of specular hematite. The blue acicular tourmaline is deficient in iron compared with the earlier brown variety (Brammall and Harwood, 1925) and points to an early decline in the iron content of the mineralising fluid during stages (i) and (ii). It is therefore considered that the deposition of hematite is a later distinct phase of mineralisation which took place after the deposition of cassiterite, from a fluid enriched in iron.

Lodes in southern and western Dartmoor provide specimens of pegmatitic aspect, which generally contain an early generation of potassium feldspar intergrown with massive dark-brown tourmaline. Cassiterite, again in association with blue acicular tourmaline and quartz, forms a second phase of mineralisation. Late red, earthy hematite and small quantities of kidney ore are more common than specular hematite, which is either absent or is a very minor constituent.

4. Distribution of mineralisation

Apart from scattered sites around the granite margin, hydrothermal mineral veins are virtually absent from the high ground of north-central Dartmoor and the extreme southern part of the pluton. There is, however, much evidence of early stream-tin working in these areas and it must be assumed that the mineralised zone has been removed by erosion. That this zone is thin, and lies immediately below the roof of the pluton is suggested by the relatively shallow depths of the workings examined, by the concentration of lodes close to the granite margin, and by the presence of metasediment as brecciated fragments in the ores from the Huntingdon Mine (SX 671 670), Hexworthy Mine (SX 656 718) and Stoneslade Tor (SX 712 783).

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The late phase of specular hematite mineralisation, which is particularly well developed in the central and north-eastern districts of Dartmoor, is considered to be the result of enrichment of the ore fluids by leaching granite contaminated by the digestion of locally iron-rich country rock.

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Hydrothermal activity at Bostraze Clay Pit (Abstract):

by N.J. Jackson and I.R. Wilson

Bostraze Clay Pit lies within a N.W. - trending (1.5 km. long x 50-100 m. wide) zone of intense argillic alteration, in the roof of the Land's End granite pluton, 1.5 km. east of St. Just. Hydrothermal activity has produced two distinct alteration assemblages: a quartz-sericite ± tourmaline assemblage, developed adjacent to quartz veins in a N.W. - trending sheeted fracture system and a quartz-kaolinite□ illite assemblage within and surrounding the sheeted vein system.

Fluid inclusion populations in samples of vein and alteration material indicate that this small area has been subject to a complex hydrothermal history. Three phases of activity can be distinguished in the homogenisation temperature ranges 400-460°C, 300-340°C and < 70°C. Mineralisation (cassiterite) and phyllic alteration appear to be associated with the first phase of activity and the argillic alteration (kaolinisation) is tentatively interpreted as being related to the last phase. A low temperature origin for the argillic alteration is supported by recent D/H and ^{18}O/ ^{16}O data.

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POTASSIUM-ARGON ISOTOPIC AGE DETERMINATIONS FROM SOME NORTH DEVON MINERAL DEPOSITS

by P.R. Ineson, J.G. Mitchell and F.J. Rottenbury

Abstract. Sixteen new K-Ar isotopic ages are reported from diverse mineral deposits in North Devon. Clay mineral concentrate fractions of clay gouge and altered wallrocks from lead-zinc, lead-silver, lead-iron, iron-manganese, copper-iron, and copper lodes are reported. The lodes are emplaced in strata which range in age from Devonian (395-345 m.a.) to Carboniferous (345-280 m.a.) K-Ar isotopic ages ranging from 364 ± 4 to 294 ± 3 m.a., are interpreted as recording thermal events in the potassium-rich clay mineral assemblage which are most probably related to metasomatism and mineralisation. The conclusions drawn from these analyses support a hypothesis of one main episode of mineralisation during upper Carboniferous times.

1. Introduction

The ore deposits of Exmoor are part of the West of England mineralised province. Particular attention has previously been focused on the Cornubian mineralisation and especially the tin-tungsten ores of south Devon and Cornwall, rather than on the diverse loads in the northern area. The close relationship between mineralisation and intrusives in the south are so well demonstrated as to make it a type locality.

The mineralisation of Exmoor has been discussed briefly by numerous authors (e.g. Carruthers *et al.*, 1915; Dewey, 1921 and 1923) but only described in detail by Dines (1956) and Rottenbury (1974). Earlier interpretations relate the mineralisation to the ore deposits in the south and attribute to it the same genetic affinities. Horizontal temperature zonation is not generally apparent however, while the shallow depth of mining precludes possible evidence for a vertical temperature zonation. The ores do, however, illustrate a relationship between a lithological control of deposition, with a

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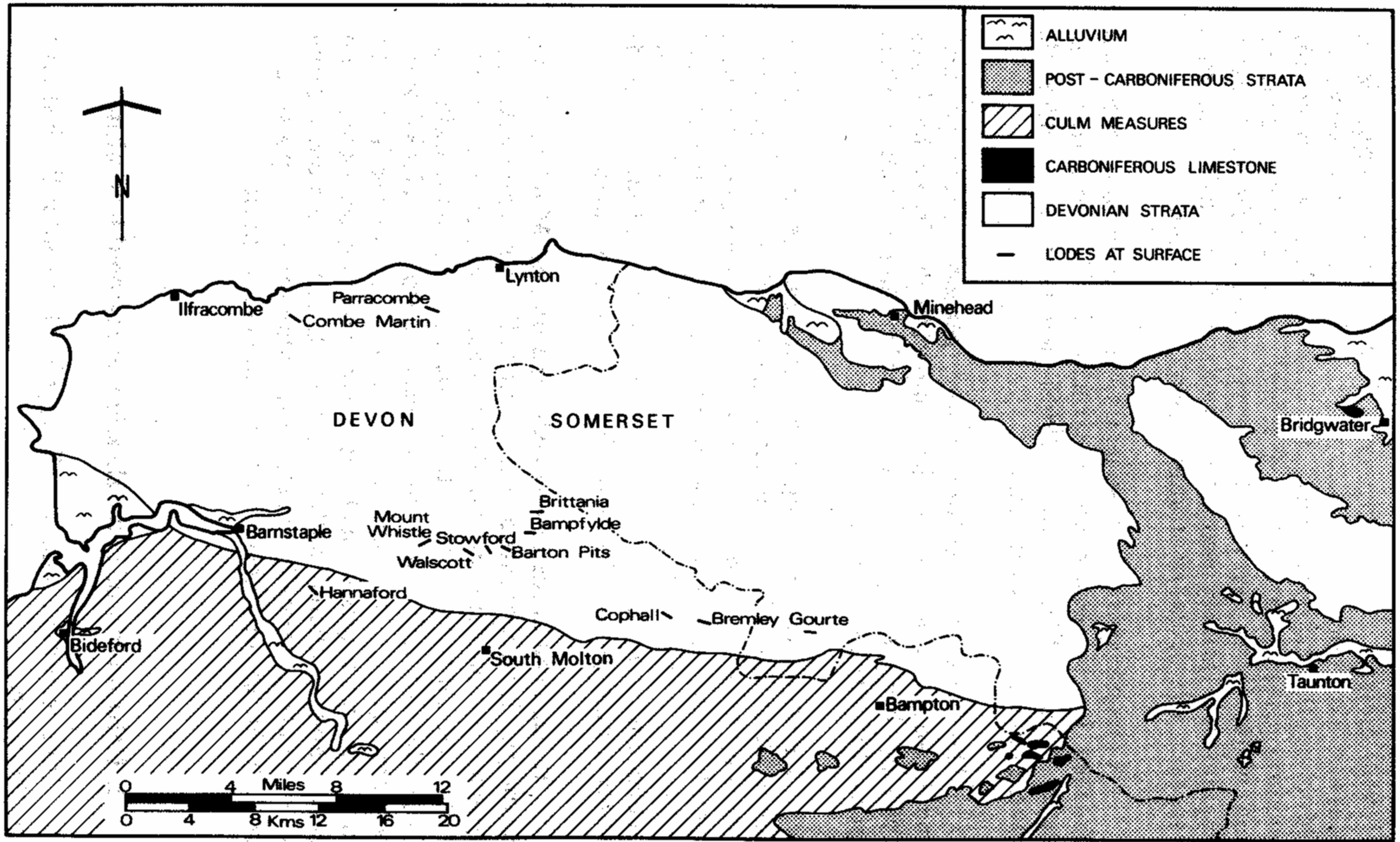

Figure 1 Sketch map of North Devon and West Somerset, with major stratigraphical units and sampled mineralised localities (after Dines, 1956).

different mineral content between arenaceous and argillaceous wallrocks. The lodes, which have a distinct linear appearance with a predominantly east-west trend, have been related to underlying structural features, rather than genetically related igneous material (Rottenbury, 1974). Likewise the absence of some of the tin-tungsten ores, almost invariably associated with the acid igneous rocks, tends to support a structural or stratigraphical interpretation.

A general synthesis of structural zones in the south-west was published by Sanderson and Dearman (1973), while structural studies (Evans, 1922; Webby, 1965) demonstrate considerable tectonic complexity in this area.

The location, mineralogy and genesis of the Exmoor mineralisation, considered to be comparatively simple by Dines (1956), has been shown to be more complex than previous workers supposed (Rottenbury, 1974).

The inconclusive geological evidence and the sparse isotopic evidence stimulated the present investigation which forms part of a continuing programme of re-appraising the isotopic ages of British mineral occurrences using the potassium argon dating method (Ineson and Mitchell, 1972, 1974a and b, 1975) to supplement the independent analyses performed using other isotopic methods (mainly uranium-lead and model lead) in the same area.

It has been argued previously (Ineson and Mitchell, 1974a) that the physico-chemical conditions prevailing when many types of epigenetic ore minerals are emplaced, being sufficiently intense to produce second generation 'metasomatic minerals' in the wallrocks, are equally of sufficient intensity to reset the isotopic 'clocks' of the new, and of any reconstituted minerals. By determining the potassium-argon ages of such secondary minerals separated from metasomatically altered wallrocks, or by dating potassium-bearing vein minerals (genetically related to the other ores) it is possible to ascertain the age of emplacement of a mineral deposit.

2. Relationship of mineralisation to regional tectonism

Five tectonic events are distinguishable within the Exmoor area, these are:

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(i) Intrusion of Late Devonian/Early Carboniferous lavas (Bittadon 'Felsite')

(ii) Late Carboniferous folding, faulting, thrusting and development of slaty cleavage about general east-west axes.

(iii) Intrusion of the Cornubian batholith to the south and volcanism in Exmoor. The minor intrusives at Hestercombe, Fremington and Rose Ash are all possibly contemporaneous.

(iv) Tertiary uplift and intrusion of the Lundy granite.

(v) Miocene 'Alpine' orogeny with the reactivation of existing fractures.

Arguments have raged as to whether the lodes fill fissures or faults. It may be considered coincident or not that the predominant strike direction for the lodes is generally about an east-west axis.

Webby (1965) sought to prove an association of the iron mineralisation with the hinges of cleavage folds, the ore shoots being formed at the junction of cleavage planes and a/c joints. Rottenbury (1974) refuted the proposal for all but minor mineral occurrences.

A rose-diagram (see Fig. 3) of predominant lode directions from the present sampled sites does illustrate a regional east-west trend with tenuous possibilities that Pb-Zn lodes may be differentiated from Cu and Cu-Fe and Fe lodes on general strike directions with an arc of 40^{o} about an east-west line. If either the general or specific lode trend is compared with that of the southern part of the south-west peninsula there is a remarkable coincidence which may reflect an overall tectonic trend and a cosanguineous mineral provenance.

3. Geological evidence for age of mineralisation

Various theories have been proposed for the age of deposition of the metalliferous lodes. Pattison (1865) suggested that the local occurrence of gold was genetically related to the greenstone intrusives of the area, while a sedimentary origin for the iron ores was proposed by Smyth (1859);who described the deposits as ferruginous nodules.

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The Tertiary age postulated by Shearman (1962) for the north-west trend, and the different mineral contents, suggested two separate periods of mineralisation. Likewise Rottenbury (1974) recognised two distinct paragenetic mineral groups which occupy fissures with different trends. While the observation may suggest different times of emplacement, no cross-cutting relationships can be observed and the point cannot be clarified.

A paragenetic study by Rottenbury (1974) resulted in the conclusion that there is no indication of an early and a late stage of mineralisation in any one lode, as neither banding of the minerals, nor rugs or fine crystallisation are evident. He therefore concluded that evidence for post-depositional movement is minimal.

In considering the time of deposition, Rottenbury (1974) could do no more than reiterate Moorbath's (1962) age determinations, but did note that the ores occurred in fissures in folded Devonian and Carboniferous strata, The mineralisation must be younger than the movements which formed the fissures, and older than the north-trending faults which cut the fissures but do not affect the New Red Sandstone of the Brendon Hills. A subsequent publication by Rottenbury and Youell (1974) suggested that deposits are of 'Permo-Triassic age as a fissure filling in the Devonian'.

Moorbath (1962) reported a personal communication from F.W. Dunning, who suggested the possibility of a pre-Hercynian age for these deposits. As no similar occurrences in the Culm (Carboniferous) rocks had been located together with the fact that the principal folding in North Devon is of Asturic (i.e. Stephanian) age, he considers it is most probable that the ores, which are strongly sheared, were emplaced prior to the main fold movements.

The geological evidence suggests two possibilities. A Devonian/early Carboniferous emplacement with subsequent deformation in the Hercynian Orogeny (late Carboniferous in the area) or a post-Carboniferous, pre-Triassic deposition with shearing of the lodes (i.e. cataclystic textures) probably during the Miocene (Alpine) Orogeny.

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TABLE 1

Reference Number	Sample Locality	Clay Mineral Concentrate Fraction	K_2O Content (%)	Radiogenic Argon Content (mm^3gm^{-1})	Atmospheric Contamination (%)	Age (m.a.)
CM1	Combe Martin Mine - Main lode Pb-Ag±Cu, Fe (s.s. 588.464)	ILLITE	1.72	$(1.82\pm0.02)\ 10^{-2}$	24.7	294±4
B4	Bampfylde Mine - North lode Cu-Fe±Ag, Au (s.s. 738. 382)	CHLORITE-KAOLINITE-Illite	0.62	$(6.51\pm0.08)\ 10^{-3}$	42.3	294±4
G1	Gourte Mine Cu-Fe (s.s. 827. 282)	Illite-Kaolinite	1.56	$(1.64\pm0.02)\ 10^{-2}$	16.2	295±3
BP1	Barton Pits Mine Fe-Mn (ss. 724.320)	ILLITE	1.82	$(1.93\pm0.02)\ 10^{-2}$	33.1	296±3
BR1	Bremley Mine Cu-Fe (s.s. 817.283)	ILLITE-Chlorite-Kaolinite	L00	$(1.06\pm0.01)\ 10^{-2}$	36.1	296±3
W1	Wallscott Mine - lode exposure Fe-Mn (s.s. 700.323)	ILLITE-Kaolinite-Chlorite	3.39	$(3.60\pm0.04)\ 10^{-2}$	15.2	296±3
S1	Stowford Mine - lode exposure Fe-Mn (s.s. 712.320)	ILLITE-Kaolinite	3.89	$(4.16\pm0.05)\ 10^{-2}$	33.5	299±3
BT1	Brittania Mine Ba-Cu-Fe (s.s. 746.335)	ILLITE	2.51	$(2.71\pm0.03)\ 10^{-2}$	20.4	301±3
MW1	Mount Whistle-Wheal Charles Fe-Mn±Cu (s.s. 686.318)	ILLITE-Chlorite	1.14	$(1.25\pm0.02)\ 10^{-2}$	25.9	306±3
CM2	Combe Martin Mine-Knap Down lode Pb-Ag±Cu, Fe (s.s. 598.467)	ILLITE	0.83	$(9.20\pm0.14)\ 10^{-3}$	34.4	308±3
WVI	Parracombe Mine-Wheal Vervale Pb-Fe (s.s. 628. 465)	Illite-Chlorite-Kaolinite	0.34	$(3.80\pm0.06)\ 10^{-3}$	62.6	309±5
B2	Bampfylde Mine - mill tailings Cu-Ft±Au. Ag (s.s. 735.344)	ILLITE-Kaolinite-Chlorite	1.53	$(1.89\pm0.02)\ 10^{-2}$	25.7	341±3
B1	Bampfylde Mine- lode Cu-Fe±Au, Ag (s.s. 735.344)	KAOLINITE-Illite-Chlorite	0.66	$(8.24\pm0.11)\ 10^{-3}$	60.4	344±4
B3	Bampfylde Mine – lode Cu-Fe±Au, Ag (s.s. 735.344)	CHLORITE-KAOLINITE-Illite	0.39	$(4.96\pm0.07)\ 10^{-3}$	49.0	348±4
COI	Cophall Mine Cu (s.s. 812.283)	ILLITE-Chlorite	0.62	$(8.45\pm0.10)\ 10^{-3}$	38.8	362±4
ECI	Hannaford Mine-East Combe Pb-Zn±Ag,Sb,Cc,Fe, (s.s. 607.297)	Illite-Chlorite	0.62	$(8.30\pm0.10)\ 10^{-3}$	38.1	364±4

Mineralogical Constituents

Major components in upper case

Minor components in lower case

Decay Constants

$$\lambda_e = 0.584 \times 10^{-10}\ yr^{-1}$$

$$\lambda_\beta = 4.72 \times 10^{-10}\ yr^{-1}$$

$$\frac{^{40}K}{K} = 1.19 \times 10^{-2} \text{ atom per cent}$$

Table 1 Potassium-argon determinations and mineral composition of sixteen samples from the mineralisation of Exmoor.

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4. Previous geochronological data

Model lead isotopic ages of the mineral deposits have been reported by Moorbath (1962), while K-Ar analyses were undertaken by Dodson and Rex (1971).

Moorbath reported three model lead isotopic ages from the Combe Martin area which indicated emplacement at 370 ± 50 m.a. and concluded that "there is little doubt that the Combe Martin mineralisation occurred possibly in late Devonian times".

Dodson and Rex reported dates of 333-288 (312 ± 32) m.a. which suggested a late Carboniferous folding episode. Roberts and Sanderson (1971) also reported 340-320 and 310-270 m.a. events and reiterated the view that the development of slaty cleavage in the northern area of the south-west England peninsula was of a late Carboniferous age.

Sanderson and Dearman (1973) noted that K-Ar ages have been widely interpreted as dating uplift and cooling rather than metamorphism, and that this is appropriate with respect to the former analyses for the 365-345 and 340-320 m.a. ages. The granites of the south-west yield K-Ar ages of 280-270 m.a. (Dodson and Rex, 1971) and cut the primary folds and some of the later folds. All the results are consistent with a view that the cooling of the batholith (280-270) is coincident with a period of uplift in north Devon (310-270). The earliest date for the primary folding in the area is 290 m.a. as it affects Westphalian rocks in Devon and lower Stephanian rocks in the Somerset Coalfield (Butcher, 1961).

5. New K-Ar geochronological data

Sixteen sample analyses from twelve mines are reported in Table 1, and Fig. 2. Details of the samples (collected by E.J.R.) are given in Table 1 and illustrated in Fig. 1. Material was obtained from within lodes, altered wallrocks, spoil heaps and mill tailings. The analytical methods have been outlined previously (Ineson and Mitchell, 1975). Ages are quoted with one standard deviation estimate of the analytical precision. Samples from several additional localities were investigated, but contained a predominance of chlorite and kaolinite and were not amenable to K-Ar analysis.

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The range of isotopic ages reported here is from 294 ± 3 to 364 ± 4 m.a. and is summarised in Fig. 2.

The data fall into two distinct groups, distinguishable on the basis of isotopic age and host rock. The younger group of ages comprise eleven of the data, all from horizons in or below the Pickwell Down sandstone; among them eight show concordant ages whose mean is 296 ± 4 m.a. The remaining three range up to 309 m.a. and may represent minor contamination by unaltered host rock material. The older group of ages comprise three samples of diverse mineralogy from Bampfylde Mine which are concordant with mean age 344 ± 4 m.a., and which lie in the upper levels of the Pickwell Down Sandstone Series. Their isotopic age is close to that of the host rock formation. The two remaining samples of the group have ages greater than the limit imposed by the host rock stratigraphy and must be discounted from further discussion of mineralisation on the grounds that they probably reflect a partially disturbed isotopic age of an altered detrital clay fraction in these upper Devonian-lower Carboniferous beds. In these, as in all other samples, contamination by non-clay silicate minerals can be discounted since XRD analyses showed no evidence of their presence.

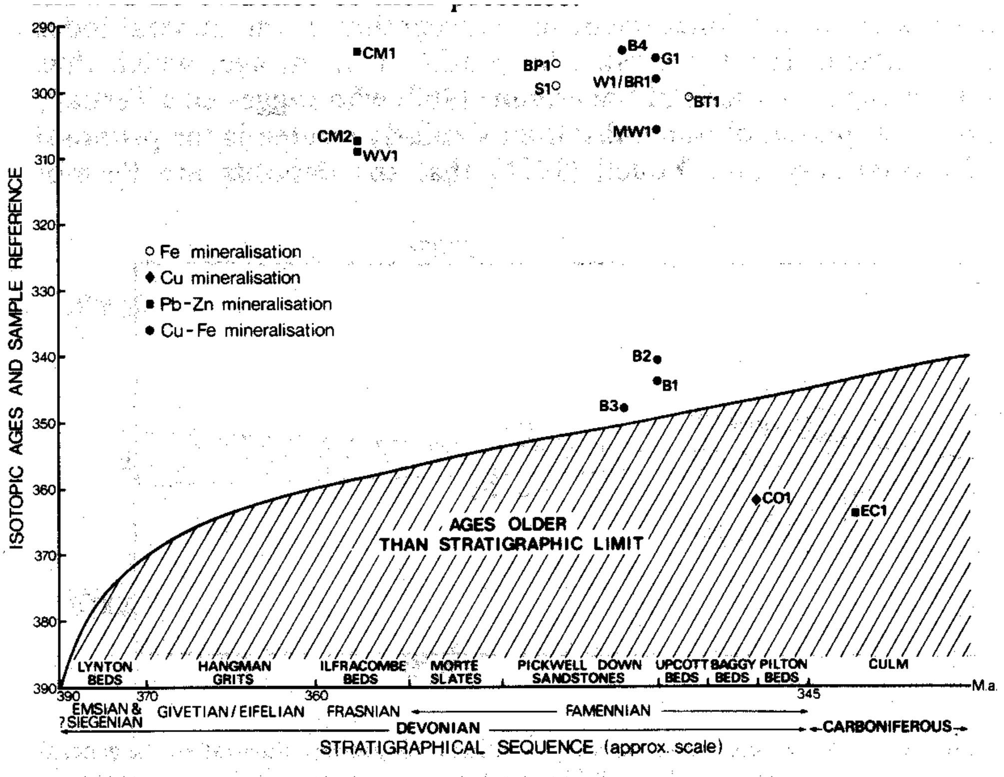

Figure 2 Compilation of isotopic age results in relation to their stratigraphical position and general mineralogy.

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6. Interpretation of K-Ar ages

The results here reported may be considered in a geochronological context and carry implications on the genesis of the North Devon ore deposits.

The inconclusive evidence from whole rock slates and mudstones dated by Dodson and Rex (1971) suggestive of a regional cleavage/uplift age of around 320 m.a. for much of the area of North Devon may be reflected in the results from Bampfylde. We cannot in any way preclude the possibility that the analytical results for this locality represent the 'regional age', however we would argue that the ages (approx. 345 m.a.) imply that the regional age must be at least this old and that the inference may be drawn that the deposits are of the same age or older than 345 m.a.

The neighbouring localities laterally within 25 km and all lying within the same vertical succession (i.e. the majority of the 296 m.a. group) cannot therefore reflect a regional age and must represent a distinct isotopic event in that area, considerably younger than the regional age (of the order of 50 m.a.) and most probably associated with emplacement or rejuvenation of the mineral lodes investigated. The lodes must be pre-296 m.a. in age, which thus refutes the hypothesis of Shearman (1962) who suggested a Tertiary proximal period of mineralisation; similarly refuted is the proposal of Rottenbury and Youell (1974) that the deposits are Permo-

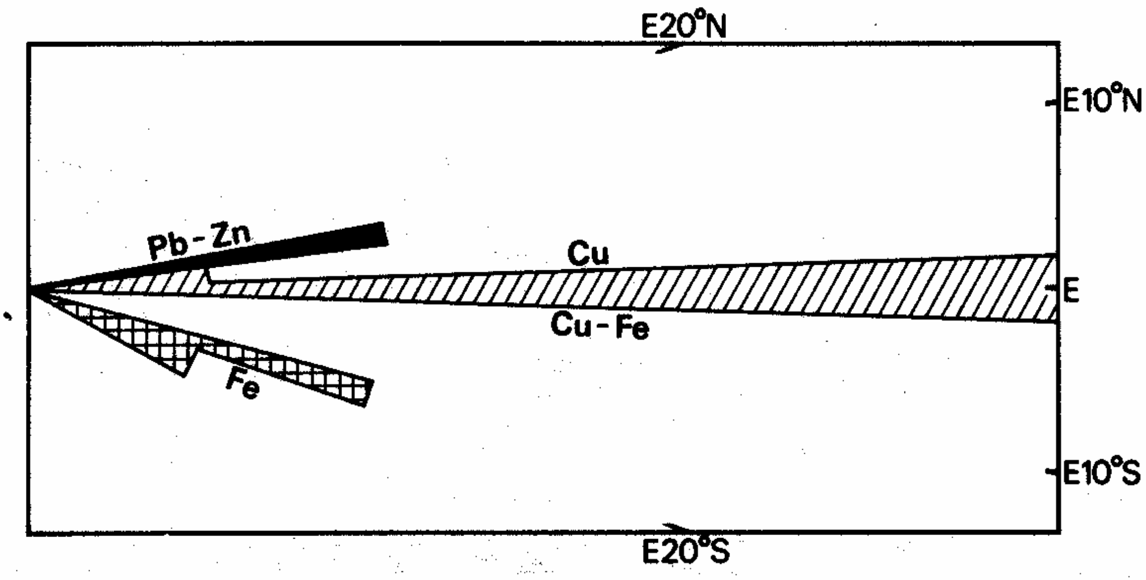

Figure 3 Rose diagram of the lode directions (16 samples) illustrating the general east-west trend, but differentiating the specific strike on mineralogical content.

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Triassic fissure fillings. The contention of Fitch and Miller (1964), that a primary mineralisation event at 225 m.a. is of widespread occurrence in the United Kingdom, cannot be advocated for this area.

In addition, on geological grounds, we believe the results are compatible with the suggestion that the genesis of North Devon base metal mineralisation is consanguineous with that of south Devon and Cornwall. The coincidence of the geochronological data, mineralogy and fault trends of the two areas suggest a close connection. Fig. 3, which illustrates the trend of the lodes analysed in this paper, is comparable with the overall trend for the south-west area, where four sets of faults are delineated as north-west, northeast, east and a minor north-south set. These are comparable with the radial and east-west faults of the southern area (see Dines, 1956); they parallel the axes of the batholith and radiate out from individual plutons. Furthermore, the mineralogy of North Devon is similar to the outer aureoles of the metalliferous mining region of the south where copper and lead with secondary minerals predominate. Rottenbury (1974) considered that, in general, the metallogeny of Exmoor is simple when compared with that of Cornwall, but that workings are shallow and that higher temperature suites could exist at depth.

The hypothetical Exmoor Thrust (Bott. et. al. 1958) with suggested northward transport of eight miles, and dipping south. occupies a similar position, with respect to the Variscan Chain. to the Faille du Midi in Belgium, which is calculated to have transported Devonian rocks 30 km northward (Fourmarier. 1933). A thrust of this magnitude or less, dipping towards the Cornish mineralised batholith, could possibly provide a link between source and deposition for the Exmoor deposits. The mineralised fluids. emanating in vertical and lateral directions from the plutonic centres. could have utilised such channelways, which may have facilitated the northerly elongation of the mineralised aureole.

7. Conclusions

The majority of the ore deposits of Exmoor were emplaced or rejuvenated inter-- or pre-- the Hercynian Orogeny, most probably at a time close to 296 m.a. This result, together with other geological factors, suggests that the genesis of the Exmoor ore deposits is consanguineous with the deposits of south Devon and Cornwall.

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The probable age of some cassiterite mineralisation in S.W. Cornwall (Abstract):

By A.N. Halliday *

K-Ar ages of muscovites, K-feldspars and albite from the Sn-Zn metallised pegmatite at Halvosso, Penryn, give evidence of argon loss in the feldspars and potassium loss in the micas. By assuming a maximum original K_2O content of 10.14 per cent for the muscovite (suggested by comparison with other S.W. England muscovite analysis) a minimum age of approximately 280 Ma can be inferred for the pegmatite. Muscovite K-Ar ages from the greisens at St. Michael's Mount and Cligga Head, combined with previous age data suggested that greisenisation occurred sometime between 280 and 260 Ma. K-Ar and ^{40}Ar-^{39}Ar studies of K-feldspars in paragenetic association with cassiterite and suphides from fissure veins and replacements yield low ages related to the structural state, and by comparison with the structural state of the Halvosso K-feldspar (dated at 280 Ma), a minimum age of 260 Ma seems likely.

K-Ar ages of basaltic dykes from the Lizard Complex (Abstract): by A.N. Halliday* and J.G. Mitchell

K-Ar ages of eleven whole rock samples from the latest phase of NW-SE trending epidorite dykes between Lowland Point and Porthoustock (E. Lizard) range between 131 and 403 Ma. The younger ages are interpreted as reflecting argon loss due to Mesozoic hydrothermal activity, and weathering. The older ages (from quarry exposures away from mineral veins), agree with those previously determined on biotites and hornblendes from surrounding schists, and their concordancy implies the timing of an "event" such as the obduction of the Complex.

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The application of the V.L.F. EM 16 prospecting technique to selected mineralised areas in south west England (Abstract):
by K. Atkinson and M.J. McCullough

Seven mineralised areas in Devon and Cornwall were surveyed. Anomalies corresponding to mineralisation were observed over the Perran Iron Lode, at Sourton Tors, Great Rock and Wheal Exmouth, extensions of the known mineralisation being indicated. Where mining contamination is severe (Tregonning-Carleen, Wheal Alfred and Cligga Head) the technique produces poorly defined anomalies. Depth estimates for the mineralised bodies agree with other geophysical and borehole data. The dip direction of orebodies is discernible from V.L.F. field profiles where the ore is a good conductor. Filtered data show overall shape and pitch of the orebodies but neither field profiles nor filtered data indicate the type of mineralisation. Pyrite, pyrrhotite, limonite and argentiferous galena give high intensity responses; ores of copper and micaceous hematite produce low intensity anomalies; sphalerite, baryte and hematite show negligible response. Water-and gouge-filled faults produce high responses comparable to pyrrhotite. Provided sufficient magnetic contrast exists geological features are recognisable, for example the granite/slate contact at Tregonning and lithological boundaries at Sourton Tors. These evaluation surveys prove the V.L.F. EM method a useful and rapid reconnaissance technique in mineral exploration and geological mapping in south west England.

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PALAEOMAGNETISM, AND THE DYNAMOTHERMAL HISTORY OF THE ST. PETER PORT GABBRO, GUERNSEY

by Ernest A. Hailwood and Christopher Garrett

Abstract. The St, Peter Port gabbro exhibits two distinct groups of palaeomagnetic stable remanence directions that are approximately antipodal with respect to each other. Gabbro samples close to the Bordeaux diorite possess a stable magnetic remanence with a shallow NW direction, whereas gabbro samples from more distal sites possess a stable remanence with a shallow to intermediate SE direction. It is concluded that the St. Peter Port gabbro may have been originally emplaced, and acquired its stable magnetisation, during a period of constant geomagnetic field polarity, and that the adjacent Bordeaux diorite was subsequently emplaced during a later period of opposite polarity. Emplacement of the diorite may have caused localised reheating and remagnetisation of the gabbro within the contact zone. A similar effect is observed near the contact of a microgabbro dyke which cross-cuts the gabbro, and these results place constraints on the possible duration of the cooling interval of the gabbro.

Studies of the magnetic susceptibility anistropy of these samples reveal well-defined magnetic foliation planes, which, at many sites, are closely parallel with visible layering within the gabbro. It is concluded that stresses developed in the gabbro during emplacement of the adjacent diorite may have caused a reorientation of the magnetic foliation planes so that they became aligned more closely parallel with the diorite contact.

The mean palaeomagnetic pole position for the St. Peter Port Gabbro differs significantly from a pole of similar age for southern Britain.

It is tentatively inferred that significant relative displacement may have occurred between the northern Channel Islands and southern Britain in post-Lower Cambrian times.

1. Introduction

The island of Guernsey is situated some 30 miles west of the Cotentin peninsula of NW France. It comprises two distinct geological parts, a southern metamorphic complex of Precambrian age, and a younger northern plutonic complex consisting of five separate major intrusions. In sequence of emplacement there are the St. Peter Port gabbro, the Bordeaux diorite, the Chouet granodiorite, the Lancresse granodiorite, and the Coho adamellite.

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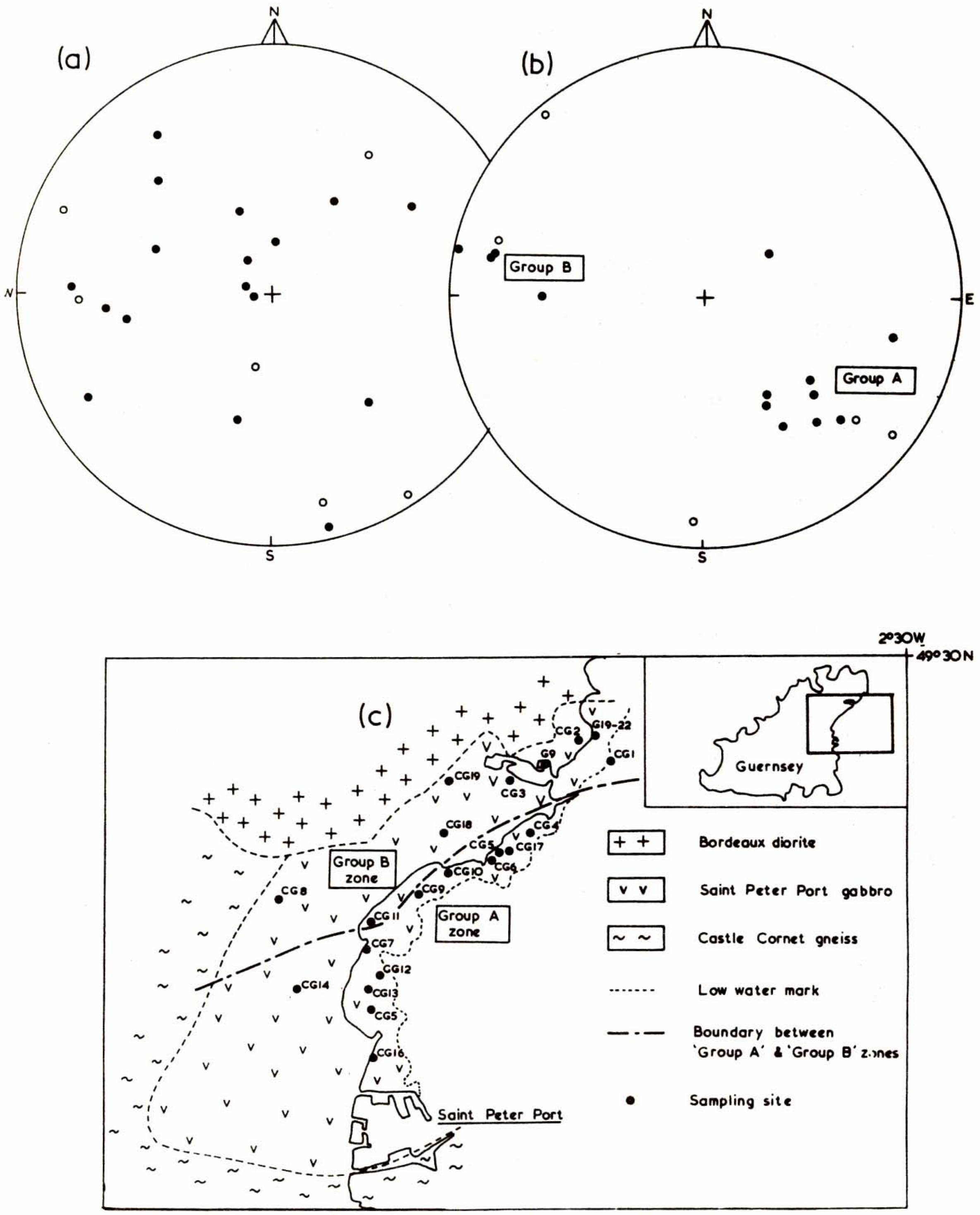

Figure 1. Polar stereographic projection of mean site directions of remanence. (a) Before laboratory demagnetisation. (b) After optimum demagnetisation in alternating fields. Solid symbols lower hemisphere, open symbols upper hemisphere. (c) Spatial distribution of sites with characteristic "Group A" and "Group B" magnetisations.

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This paper describes a preliminary study of the magnetic remanence and susceptibility anistropy of the oldest of these intrusions, the St. Peter Port gabbro. The study was undertaken to elucidate the mode of emplacement and cooling history of the gabbro, and to establish the position of the Channel Islands with respect to other parts of the European craton in Early Palaeozoic times.

The only available radiometric age determination for the Guernsey plutonic complex is from the youngest intrusion, the Cobo adamellite, which has yielded a Rb/Sr isochron of 570 ± 15 m.y. (Bishop et al, 1975). However emplacement of all five bodies is believed to have taken place at a late stage in the Cadomian orogeny, and on these grounds the age of the St. Peter Port gabbro is thought to be early Cambrian.

A well-developed layered structure occurs throughout the central part of the gabbro, probably as a result of fractional crystallisation and gravitational settling of crystals within a magma chamber. The thickness of individual layers ranges from a few centimetres to hundreds of metres, and they commonly dip at a low angle to the southwest.

An average of six separately orientated specimens was collected at each of 19 sites, distributed as far as possible over the whole outcrop of the gabbro. Because inland exposures are poor, however, there is a tendency for sites to be concentrated along the well-exposed coastal sections (Fig. I c). Most samples were collected by means of a portable coring drill, which allowed easy penetration though the weathered surface layer, and wherever possible orientation was performed by sun compass rather than magnetic compass.

2. Palaeomagnetic results

Palaeomagnetic measurements were performed on a Digico computerised slow-speed spinner magnetometer (Molyneux, 1971) and the remanence directions of the six separate samples at each site were combined into site-mean directions to average out local inhomogeneities and small orientation errors. The mean site

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remanence directions are widely scattered over both hemispheres before laboratory treatment of the samples (Fig. la), but after alternating field (A.F.) demagnetisation in peak fields of 100 to 300 Oe there is a clear polarisation of the directions into two groups that are approximately antipodal with respect to each other, designated "Group A' and "Group B" (Fig. lb). This distribution indicates that the cooling history of the gabbro must have spanned at least one reversal of the geomagnetic field. Examination of the spatial distribution of these two groups of remanence directions (Fig. lc) reveals that all sites within Group B (NW declination) are situated closer to the contact with the Bordeaux diorite than those in Group A (SE declination). Furthermore, a line 'drawn between these two groups follows Closely the shape of the boundary between the gabbro and diorite.

The most likely explanation for this distribution is that the gabbro was originally magnetised with a single polarity throughout, corresponding with the Group A direction, and that subsequent' intrusion of the nearby Bordeaux diorite caused local reheating within a zone approximately 1/2 km wide, parallel with the diorite contact. The intrusion of the diorite must have occurred after the geomagnetic field had reversed polarity, so that the sites within the contact zone were remagnetised with the opposite polarity to those of Group A.

This conclusion is supported by the behaviour of representative samples during laboratory progressive A.F. demagnetisation experiments. Fig. 2a shows the response of several gabbro samples from within the Group A zone, but situated close to the junction with the Group B zone. The directions of magnetisation before treatment lie close to the characteristic Group B direction, but during demagnetisation they move progressively towards the Group A direction. This behaviour suggests that the samples suffered a mild degree of reheating, sufficient to impress a moderately stable secondary (Group B) direction of magnetisation on them. However, the more-stable Group A direction was not completely eradicated, so that it could be successfully retrieved by laboratory removal of the Group B component.

Fig. 2b illustrates the response to A.F. demagnetisation of samples from the microgabbro dyke that intrudes the gabbro at site CG 17, within the Group A zone. Samples 17.1A and 17.1B are from the dyke itself, and exhibit a characteristic Group B direction.

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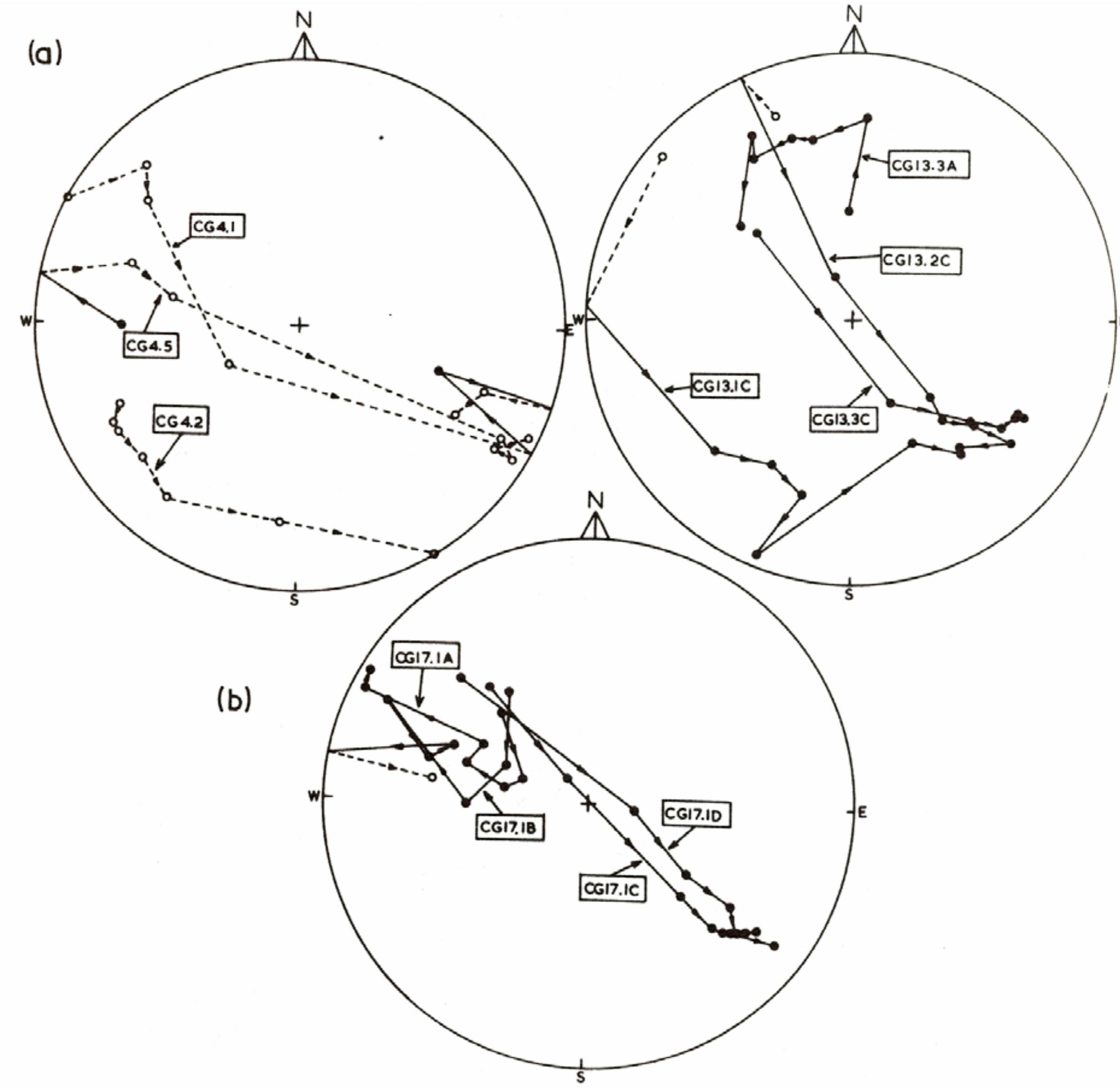

Figure 2. Response of representative samples to progressive alternating field demagnetisation (Polar stereographic projections). (a) "Group A" gabbro samples. (b) Samples from the microgabbro dyke at site CG 17, and baked gabbro contact. Successive points represent remanence directions after treatment at 50 Oersted increments, up, to 300 Oersteds.

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However samples 17.1C and 17.1D are from the adjacent gabbro, within the narrow baked contact zone of the dyke. Although these samples appear to have been remagnetised as a result of intrusion of the dyke, progressive AF demagnetisation removes this less-stable secondary component, and once again isolates the characteristic Group A direction.

These results indicate that the original emplacement of the gabbro, and the later emplacement of the diorite and microgabbro dyke must have been separated by an interval of time at least as long as the duration of a typical geomagnetic field polarity transition. This is commonly believed to be of the order of 10n to 105 years. However the possibility of several polarity transitions having taken place between the two major episodes of intrusion cannot be ruled out, so that the above may be taken as a minimum figure only. In the same way, it is not certain that the diorite and microgabbro dyke were emplaced during the same polarity interval. Nonetheless, the results illustrate the potential value of palaeomagnetic studies in attempting to unravel the history of successive intrusions and reheatings in a complex plutonic region.

3. Magnetic anistropy results

The magnetic susceptibility anistropy of the gabbro samples was measured by means of a low field torsion magnetometer (King and Rees,1962.) This property reflects the preferred alignment of magnetic mineral grains, and can be used to specify (i) the orientation of the magnetic foliation plane, and (ii) the direction of the magnetic lineation axis, within this foliation plane.

Figure 3a shows a stereographic plot of the poles to the magnetic foliation planes (triangles) and the poles to the visible layering of the gabbro (squares) where the latter was measurable. On the whole the magnetic and visible foliation planes appear to be closely parallel, so that magnetic anistropy measurements may be used to provide a rapid, accurate and convenient means of specifying the orientation of layering within the gabbro.

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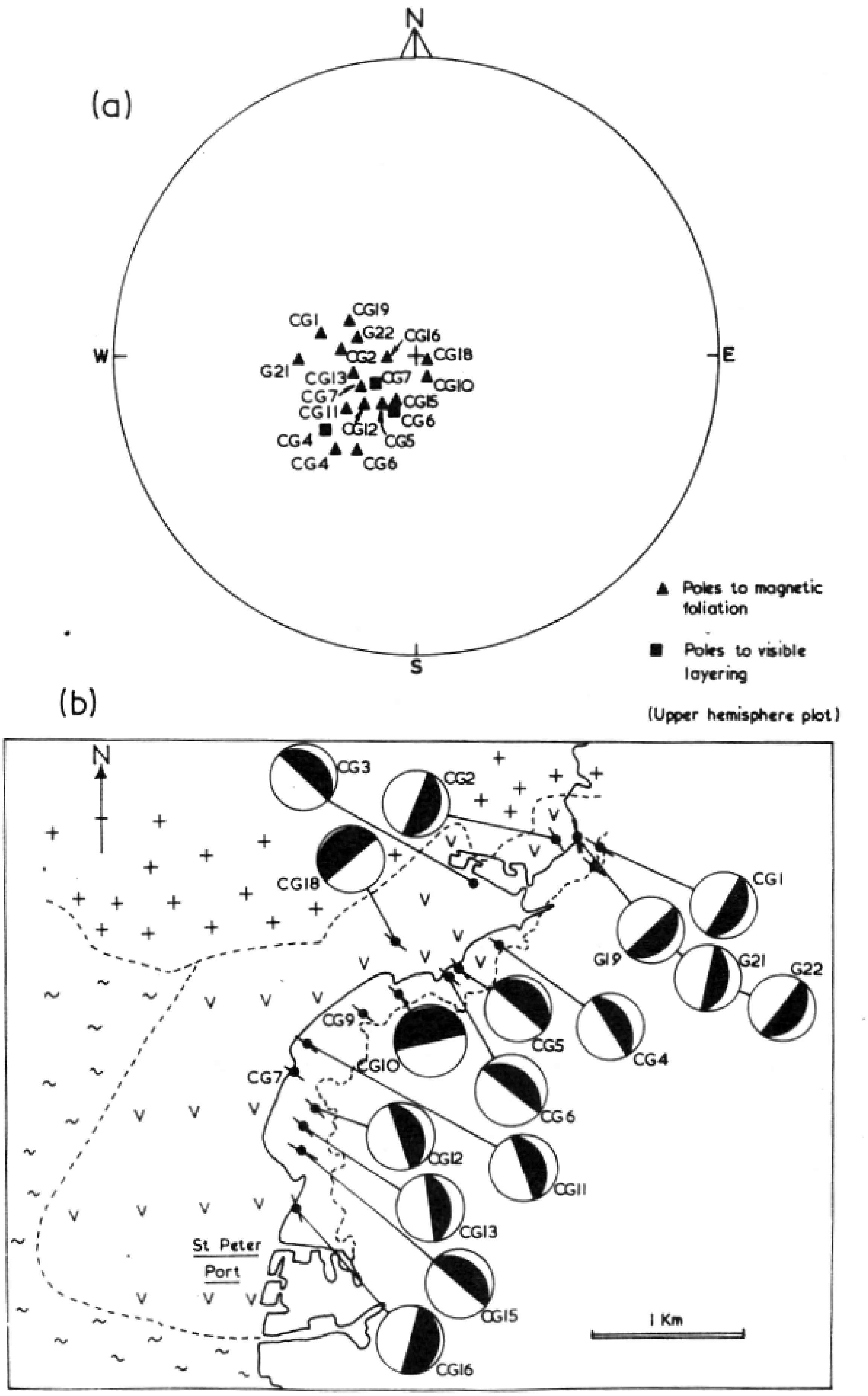

Figure 3. (a) Comparison between poles to magnetic foliation planes and visible lamination in the St. Peter Port gabbro. (b) Orientation of magnetic foliation planes and lineations (see text for explanation).

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Figure 3b illustrates the spatial distribution of magnetic foliation plane orientations at different sites. Each 'balloon' represents a stereographic plot of the mean foliation plane orientation at that site, with the upper half of the foliation plane shaded black. The main group of sites situated along the shore between St. Sampson and St. Peter's Port, and distal from the diorite contact, show magnetic foliation planes with a general dip to the southwest. Only two sites, CG10 and CG 16, depart from this trend, and may reflect local small-scale variability within the gabbro. The overall south-westerly inclination may represent either a primary dip of the gabbro layering, similar to that observed in laccolithic intrusions such as the Skaergaard complex, or a secondary tilt of the whole gabbro mass.

Conversely the five sites that are more proximal to the diorite contact (CG 1, CG 2, G 19, G 21, and G 22, Fig. 3b) show a general westerly or northwesterly dip, which is more closely parallel with the general trend of the contact. It is possible that these W to NW dipping foliation planes were developed as a result of stresses generated in the gabbro during emplacement of the diorite. The magnetic foliation planes at sites CG 3 and CG 18 differ from this trend, but are not well defined, due to a significantly higher degree of within-site scatter than at the other sites.

The orientation of the magnetic lineations is depicted by the short lines at each site location in Figure 3b, and there appears to be a consistent NW-SE trend to these lineations. It seems highly probable that these lineations are related to the direction of flow of the magma during intrusion of the gabbro, and more-detailed magnetic anistropy studies are in progress to further elucidate the dynamics of emplacement of this gabbro body.

4. Pole positions and plate tectonic implications

The overall mean remanence direction for the sixteen reliable sites is given in Table I, both before and after application of a tilt correction based on the average dip of the magnetic foliation planes. Relevant statistical parameters are also listed, together with the corresponding palaeomagnetic pole positions. These pole positions are plotted in Fig. 4, together with Palaeozoic polar wander curve for the 'stable' cratonic part of NW Europe (after McElhinny, 1974).

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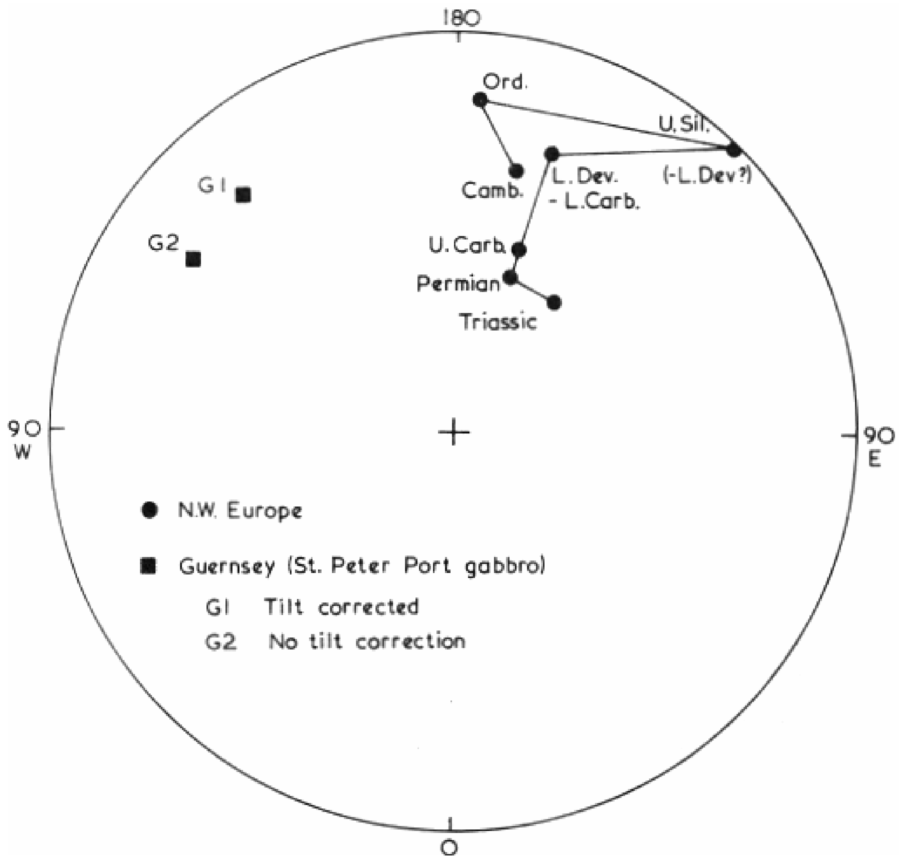

Figure 4. Comparison between palaeomagnetic polar wander path for "stable" NW Europe and Lower Cambrian poles from the St. Peter Port gabbro.

As discussed above, the St. Peter Port gabbro is considered to have been emplaced in Early Cambrian times. However, there is a clear discrepancy between the corresponding palaeomagnetic pole positions (both before and after tilt correction) and the Cambrian pole for 'stable' NW Europe derived from the Caerfai series of South Wales (Briden et al, 1970). It is tentatively inferred that significant relative 'displacement has taken place between Guernsey and southern Britain in post-Lower Cambrian time. It is interesting to note that these results are consistent with an anticlockwise rotation of Guernsey through some 500 relative to the European plate. However if Guernsey had acted as a simple 'roller block' during the oblique interaction of the African and Euramerican plates which culminated in the Hercynian orogeny, a clockwise motion would be anticipated. It must be concluded that local tectonic displacements of small blocks within this collision belt may have been extremely complex.

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Table I Mean Pole Positions and Remanence Directions

	Pole Position		Remanence Direction				
	Lat.	Long.	Dec.	Inc.	N	k	$\alpha 95$
No tilt correction	14N	237E	122.2	15	16	9.7	11.2
Tilt correction based on average dip of magnetic foliation planes	13N	224E	132.2	28.7			

N - Number of sites, k - Fisher precision parameter, □95 - radius of 95% confidence circle

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Geological results of recent deep seismic investigations in the Bristol Channel area (Abstract):

by M. Brooks.

Seismic investigations by the University College of Swansea have been carried out since 1970. The first phase of investigations comprised 30 sonobuoy refraction lines up to 17 km in length. Together with complementary seismic profiling and bottom sampling surveys, these lines provide estimates of Mesozoic thickness in various parts of the Channel. Three main synclinal areas have been defined, each containing a substantial thickness of Mesozoic strata: the Helwick syncline (containing about 700 m) southwest of Gower, the Bristol Channel syncline (about 2 km) north of the North Devon coast, and the westward extension of the Glastonbury syncline (nearly 1 km) under Bridgwater Bay. These three folds of pre-Upper Cretaceous age largely define the present day shape of the Bristol Channel. Faulting is of secondary importance but a major strike fault, the Central Bristol Channel fault, affects the northern limb of the Bristol Channel syncline and has a post--Kimmeridgian southerly downthrow of I km in places. Refractor velocities obtained along the various seismic lines permit speculation on the geology of the Palaeozoic floor of the central and inner Bristol Channel area, and a Lower Palaeozoic inlier in mid-Channel is tentatively suggested.

Recent long seismic lines based on quarry blast recordings in South Wales. and land recordings of large Aquaflex charges detonated in the Bristol Channel provide information on the deeper structure, including the disposition of the Precambrian basement. Provisional geological interpretations are discussed, and the case for a major thrust under Exmoor is re-examined in the light of the new seismic data.

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THE GEOCHEMISTRY OF SOME BIOTITES FROM THE DARTMOOR GRANITE

by S. Al-Saleh, R. Fuge, and W.J. Rea

Abstract. Samples of biotite have been separated from the Marginal granite (2). Big Feldspar granite (5), Poorly Megacrystic granite (3) and the 'Aplitic' granites (4) of Dartmoor. Biotites from the 'Aplitic' granites are similar in major element chemistry to those of the Big Feldspar granite. Biotites from the Poorly Megacrystic granite are richer in Si, Al, Mn, Li, and F than are those from the Big Feldspar granite, while Fe^{2+}, Ti, Mg and H_2O are highest in the biotites of the Big Feldspar granite. Appreciable quantities of phosphorus occur in all of the biotites but highest values occur in the Big Feldspar granites. Lead is enriched in all of the biotites (in excess of 200 ppm on average), but is richest in those of the 'Aplitic' granites. Cr and V are richest in the biotites of the Big Feldspar granites as is C1, while Zn is highest in the biotites of the poorly Megacrystic granite.

The chemistries of the biotites of the Big Feldspar granite compared with those of the Poorly Megacrystic granite are as might be expected from the more evolved nature of the Poorly Megacrystic granite. However, it is not possible in this study to suggest how much the biotite chemistry has been derived and modified by assimilation and metasomatism.

1. Introduction

The Dartmoor pluton is composed essentially of two main granite types termed by Hawkes (1968) the Big Feldspar granite, which is equivalent to the 'Giant' granite of Brammall and Harwood (1923), and the Poorly Megacrystic granite, which is equivalent to the 'Blue' granite of Brammall and Harwood (1923). The Big Feldspar granite overlies the Poorly Megacrystic granite and has been interpreted as a marginal facies of it (Hawkes and Chaperlin, 1966).

In addition to these two main types there are also some minor rock types (Hawkes 1968). Of these, one of the most interesting is a fine-grained leucocratic granite of aplitic appearance which occurs at the contacts and as sheets, veins and Pods throughout the pluton

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but mainly in the Big Feldspar granite. Due to its appearance this granite was termed the 'Aplitic' granite by Brammall and Harwood (1923) and the term is retained in this work. Hawkes (1968) has suggested the 'Aplitic' granites represent either a re-distributed chilled margin or granitised Carboniferous sandstone.

The work in this paper represents a preliminary report of what is intended to be a detailed study of all of the mineral phases of the Dartmoor granites. The samples are from the following rock types:-

2 from marginal granites (Finer-grained variant of the Big Feldspar granite).
5 from the Big Feldspar granite
3 from the Poorly Megacrystic granite
4 from 'Aplitic' granites

Several kilograms of the granites were collected and samples for analysis and biotite separation were selected so as to be as free from weathered material as possible. The biotite samples were separated from 60-100 mesh powder using a Frantz isodynamic magnetic separator. The resultant concentrate which was about 75 to 90% biotite was hand picked and about a gram of sample was separated for analysis. Petrological work on the host rock samples revealed that some of the biotites contain inclusions of magnetite, sphene, tourmaline, apatite and zircon. However, it is felt that few of the biotites contain appreciable quantities of these inclusions (e.g. it is highly unlikely that much apatite is present as Ca cannot be detected in over half of the biotites analysed). However, several of the biotites showed some patches of chlorite which it is felt would not be removed using the present method of separation. It is estimated that the mineral samples used were more than 95% pure.

Analytical work was carried out on the host rocks and on the biotite separates using automated photometric analysis with a Technicon AutoAnalyzer (For Si, Al, Ti, Fe, P, F, Cl, Cr, and V), Atomic Absorption spectrophotometry using a Pye Unicam SP 1900 (For Mg, Ca, Mn, Li, Co, Ni, Cu, Pb and Zn) and Flame Emission Spectrophotometry also using the SP 1900 (For Na and K), while water was determined using a modified Penfield method. The analytical data for the biotites were standardised against the C.R.P.G. reference sample iron mica.

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Modal analysis of the two main granites and of the aplitic granite shows that the Big Feldspar granite has a very variable biotite content but generally contains in the region of 10-12% biotite and commonly more. The Poorly Megacrystic granite contains between 5 and 8 % biotite while the 'Aplitic' granite generally contains less than 5% biotite.

TABLE 1. **Average Chemical Composition of Granites**

	Big Feldspar Granite		Poorly Megacrystic Granite		'Aplitic' Granite
	Present Study	Survey Memoir* (1968)	Present Study	Survey Memoir* (1968)	Present Study
SiO_2	72.3	71.56	74.6	74.77	75.7
Al_2O_3	13.6	14.08	12.7	13.33	12.1
Fe_2O_3	1.17	0.55	0.67	0.44	0.63
FeO	1.74	2.21	1.11	1.38	0.43
MnO	0.06	0.06	0.05	0.08	0.03
MgO	0.5	0.6	0.22	0.47	0.08
CaO	0.92	1.55	0.5	0.63	0.34
Na_2O	2.91	3.01	3.24	2.87	2.79
K_2O	5.49	4.87	4.79	4.75	5.33
Li_2O	0.06	0.1	-	0.03	-
TiO_2	0.4	0.42	0.17	0.2	0.08
P_2O_5	0.19	0.21	0.2	0.18	0.08
H_2O	1.23	0.88	0.95	0.9	0.86
TOTAL	100.58	100	99.3	100	98.48
Trace elements in ppm					
F	1332		1887		445
Cl	620		400		275
Co	57		87		68
Ni	12		10		14
Cu	10		13		10
Pb	Tr		Tr		Tr
Zn	81		48		131
Cr	n.d.		n.d.		n.d.
V	20		5		6.5

*Based essentially on data of Brammall and Harwood (1932)

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2. The general chemistry of the granites

The chemistry of the Dartmoor granites was investigated by Brammall and Harwood (1932). The data obtained by the present authors are included along with the data obtained by Brammall and Harwood in Table 1.

Major differences in composition between the two main types are that Si, Na, Li and F are higher in the Poorly Megacrystic granite, while Al, Fe^{+3} and Fe^{+2}, Mg, Ca, K, Ti, C1, Zn and V are higher in the Big Feldspar granite. These differences are generally similar to those found by Brammall and Harwood. It has been suggested that these variations are due in part to contamination of the outer granite by assimilation of argillaceous material and in part also to differentiation from the Big Feldspar granite to the more evolved Poorly Megacrystic granite (Brammall and Harwood, 1932; Exley and Stone, 1964; Hawkes, 1968).

The 'Aplitic' granite is generally low in the basic components and is high in SiO_2. However, the composition of the aplitic granites is very variable.

3. General chemistry of the biotites

The analytical data for the biotites are presented in Table 2. The totals for the biotites which are low can partly be accounted for by non-determination of such elements as Rb, and Cs. The calculated structural formulae for the biotites (see Table 2) shows that the octahedral atoms are slightly lower than might be anticipated. Nockolds (1947) has suggested that the value for octahedral atoms in biotite co-existing with muscovite in calc-alkaline rocks should be 5.45. The same author has suggested that biotites co-existing with topaz would have octahedral totals in the region of 5.08. It is possible that the presence of tourmaline in the granites could result in low octahedral element totals in the Dartmoor biotites. However, it may also be that some undetermined divalent element(s) may be present in the biotites. It is hoped that future probe work may shed some light on this problem.

Our values for the Dartmoor biotites differ markedly from those quoted by Brammall and Harwood (1932) Particularly with regard to Si, Al, Ti, Fe and Li.

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TABLE 2. Average Chemical Analyses for Dartmoor Biotites

	Big feldspar Granite	Poorly megacrystic	'Aplitic' Granite
SiO_2	36.43	38.64	37.25
Al_2O_3	16.0	21.90	16.7
Fe_2O_3	6.13	3.62	6.92
FeO	16.1	13.38	16.14
MnO	0.33	0.55	0.56
MgO	4.66	1.93	2.54
CaO	0.16	-	Tr
Na_2O	0.40	0.44	0.34
K_2O	9.71	10.47	9.88
Li_2O	0.372	0.993	0.465
TiO_2	3.58	2.13	2.72
P_2O_5	0.29	0.137	0.13
H_2O	3.69	2.64	3.51
F	0.937	1.759	1.097
Cl	0.413	0.177	0.549
Total	99.202	98.766	98.801
Less O = Cl; F	0.488	0.781	0.586
TOTAL	98.714	97.985	98.215
Trace elements (ppm)			
Co	56	53	55
Ni	28.5	22.5	16
Cu	49	48	58
Pb	211	193	257.5
Zn	476	577	628
Cr	37	18	13
V	139	55	49

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Plots of the data on triangular diagrams, using the points Fe_20_3 + TiO_2, MgO, and FeO + MnO (Heinrich, 1946), show that the biotites fall within the granite field as defined by Engel and Engel (1960). The samples are all iron biotites but whereas the samples from the Big Feldspar granite and some of the 'Aplitic' granites fall within the field of granite micas in the triangle Mg: Al + Fe^{3+} + Ti: Fe^{2+} Mn, samples from the central granite lie outside this field; the biotites of the central granite being aluminium rich.

The major element chemistries of the biotites from the 'Aplitic' granites are similar to those of the Big Feldspar biotites, but it should be noted that the analysed biotites were all separated from 'Aplitic' granites which occur within the outer granite.

The trace element content of the 'Aplitic' granites shows that the low Cr and V values are similar to those in the poorly megacrystic granite, but lead and zinc values are extremely high.

Lead is much higher in all of the analysed biotites than for those previously reported in the literature. The highest values recorded by Wedepohl (1972) are in general less than 100 ppm. The greatest values in the present study are all much higher, with the highest from the 'Aplitic' and Big Feldspar granites. This may in part be linked with the presence of associated mineralisation and in part with the late stage potash metasomatism. Lead in potash feldspar replaces potassium and consequently could possibly replace potassium in the biotites. Bradshaw (1967) working on South-West England granites, including Dartmoor, found higher lead values in biotites and feldspars from granites associated with mineralisation. However, Bradshaw's mean value for biotites in these mineralised areas is only 31 ppm and 20 ppm for non-mineralised areas.

The lead values recorded here do not differ markedly from some of those quoted for Portuguese Hercynian granites by Neiva (1976). Again, it should be noted that some of our samples were collected from areas where there is some uranium mineralisation and it is possible that some of the lead is derived from radioactive decay of uranium.

In contrast to lead, the values recorded for zinc, nickel, manganese and copper are similar to those quoted by Bradshaw.

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A comparison of the data for the Big Feldspar granite and the Poorly Megacrystic granite reveals that generally Si, Al, Mn, Li and F increase from the Big Feldspar granite to the Poorly Megacrystic granite. This is particularly marked in the case of lithium. Ferrous iron, Ti, Mg and H^20 show decreases in concentration from the Big Feldspar granite to the Poorly Megacrystic granite. The relative composition of the two main granite types show the trends which have been observed by other workers (Neiva, 1976), in differentiated granitic bodies. However, the values recorded for ferric iron show a decrease with increasing silica content. According to Deer, Howie and Zussman (1962) and more recently Neiva, during differentiation, biotites show increasing ferric iron with increasing silica in the host rock.

Phosphorus values are not normally quoted for biotites. However, our values show that P_2O_5 is quite high in all of the biotites. There is little or no calcium in these biotites so that the phosphorus cannot be due to apatite impurities. One would expect therefore that phosphorus is replacing silicon in the biotite structure. The exact significance of the higher P_20_5 content of biotites of the Big Feldspar granite is not clear. However, Henderson (1968) found appreciable quantities of phosphorus to enter early formed minerals in the basic and ultra basic intrusions of Skaergaard, Bushveld and Rhum. It is possible that the higher charge density of phosphorus partially explains its greater incorporation in the earlier formed biotites.

The behaviour of the trace elements is in keeping with the idea of differentiation. Cr and V decrease markedly, Ni decreases somewhat and Zn increases markedly. The trend for zinc is in agreement with that observed by Viswanathan (1973) for biotites in differentiated granitic bodies. This author found that Zn/Fe ratios increase during differentiation, a trend which also shows up in the Dartmoor samples. However, the trend of decreasing Pb/Zn ratios observed by Viswanathan is opposite to that observed during the present study.

4. Fluorine and chlorine in the biotites

Both fluorine and chlorine are relatively high in the Dartmoor granite (Fuge and Power, 1969). It is apparent from the values quoted in Table 2 that substantial quantities of both elements occur in

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Structural Formulae for the average chemical composition of biotites on the basis of 24 (O+OH + F + Cl)

	Big Feldspar Granite		Poorly Megacrystic Granite		'Aplitic' Granite	
Si	5.593		5.628		5.761	
P	0.037	Z=8.000	0.018	Z=8.000	0.017	Z=8.000
Al	2.370		2.354		2.222	
Al	0.522		1.403		0.820	
Fe^{+3}	0.708		0.397		0.804	
Fe^{+2}	2.066	Y=5.048	1.628	Y=4.730	2.086	Y=4.974
Mn	0.043		0.068		0.073	
Mg	1.066		0.419		0.585	
Li	0.230		0.581		0.290	
Ti	0.413		0.234		0.316	
Ca	0.027		-		-	
Na	0.120	X=2.048	0.124	X=2.069	0.102	X-2.050
	1.901		1.945		1.948	
OH	3.775		2.562		3.618	
F	0.454	4.337	0.810	3.416	0.536	4.298
Cl	0.108		0.044		0.144	

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biotites. Whereas some chlorine may occur in fluid and soluble solid inclusions, it seems likely that most of the fluorine and chlorine occur within the biotite lattice where they are replacing the hydroxyl group.

Stormer and Carmichael (1971) used simple thermodynamic data for binary compounds of various metals with hydroxyl and the halogens to predict the behaviour of the two halogens in the igneous environment. They predicted that fluorine would occupy hydroxyl sites in preference to chlorine and only in the virtual absence of fluorine would chlorine be admitted. On the basis of these calculations little or no chlorine should have been admitted into the hydroxy minerals of the Dartmoor granite. However, the analyses reveal that large quantities of chlorine are present in the biotites.

If one considers the experimental data obtained for the behaviour of fluorine and chlorine in silicate melts of granitic composition (Koster Van Groos and Wyllie, 1968;1969,Wyllie and Tuttle, 1961;1964), it is apparent that whereas fluorine has a high solubility in the melt chlorine has a low solubility. Therefore, in the original Dartmoor melt there would have been relatively high concentrations of both fluorine and chlorine, chlorine would tend to be rejected by melt. In a granitic melt where few hydroxy minerals are being precipitated the rejected chlorine would tend to form an immiscible fluid phase in association with water and alkali metals. However, in the Dartmoor granite large quantities of biotite were precipitated and provided many potential lattice sites for fluorine and chlorine. However, as fluorine has a high affinity for the melt there is less competition for chlorine and it can occupy a relatively large number of lattice sites.

If, then, the transition from Big Feldspar granite to Poorly Megacrystic granite is due to magmatic differentiation, during the course of this differentiation the fluorine content of the biotites has increased while the chlorine content has decreased. This behaviour would be expected as the fluorine content of the melt would build up during differentiation and consequently more would enter crystallising biotite so excluding more and more chlorine (Fuge, 1977).

5. Concluding remarks

It is apparent from the data that lead is high in the Dartmoor biotites. Phosphorus content is also appreciable and varies

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significantly between the two main granite types. Fluorine and chlorine are high in Dartmoor biotites but vary greatly in the biotites of the two main types. Whether the variation in phosphorus, fluorine and chlorine in the biotites of the two main granite types is due entirely or in part to differentiation is not certain. However, much of the chemistries of the biotites from the two main granites is consistent with the idea of differentiation. It could equally be argued that many of the chemical differences are also consistent with the idea of contamination of the outer granite with assimilation of argillaceous material.

It is hoped that the continuation of this project will shed some light on the relative importance of differentiation, assimilation and metasomatism in the development of the Dartmoor granites. In particular electron microprobe analyses of mineral phases coexisting with biotite in the granites will allow a more exact assessment of the extent to which differentiation may have controlled the observed chemical trends in the granites.

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THE OCCURRENCE OF TOPAZ-RICH GREISENS IN ST. MICHAEL'S MOUNT, CORNWALL

by Farid Moore

1. Introduction

Saint Michael's Mount granite is a small stock situated in Mount's Bay immediately to the south of Marazion, Cornwall (SW515 298). The stock rises approximately 60m above sea level and at low tide is connected to the mainland by a causeway.

Field evidence reveals that intrusion at a high level in the crust has taken place by passive stoping, together with an element of forceful emplacement under the influence of the Land's End pluton stress system. The granite is in contact with slates and hornfels of the Mylor series and the contact can be seen on the east and west foreshores.

Lithologically the stock consists essentially of several varieties of porphyritic granite and granite porphyry, with local pegmatitic, aplitic and leucocratic facies. During its contractional cooling the stock suffered extensive sheet fracturing and developed two major sets of joints striking 070 and 050 and some minor jointing on the north west flank.

Mineralised, greisen-bordered veins occur within the granite. They are sub-parallel and almost vertical, being predominantly confined to the main sets of joints. Occasionally they continue into the slates where they produce no apparent alteration.

2. Petrographic notes

Megascopic K-feldspars, tourmaline and quartz crystals are set in an aphanitic groundmass of quartz, orthoclase microperthite, sodic plagioclase, muscovite and tourmaline, with accessory topaz, biotite, apatite, cassiterite, magnetite and, in rare cases, zircon.

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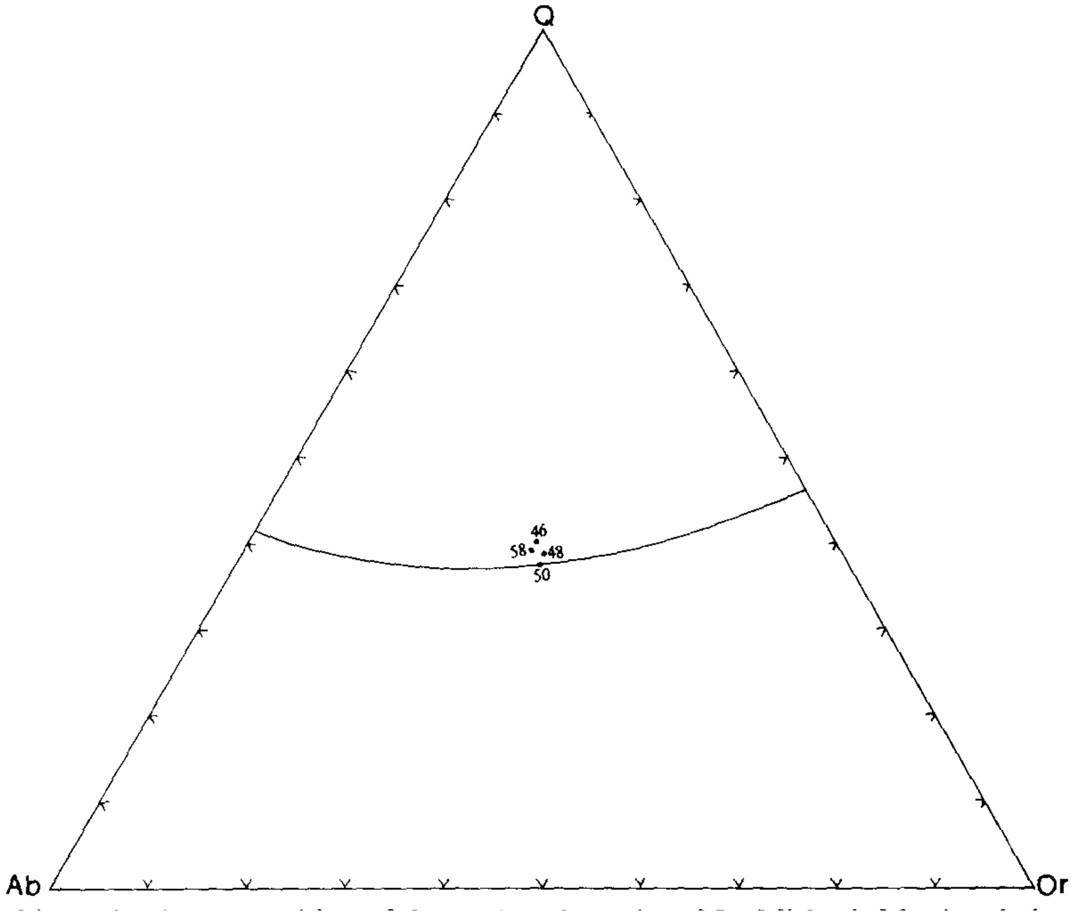

Figure 1. The composition of the unaltered granite of St. Michaels Mnt in relation to the system Q-Or-Ab-H2O (quartz-feldspar boundary and ternary minimum at 1000 bars water pressure after Tuttle & Bowen 1958).

The amount of groundmass varies between 5 to 30% and so produces several varieties of granite porphyry and porphyritic granite. Almost all of these varieties have the same mineralogical composition and there is no indication of post-magmatic alteration, however all are rich in tourmaline and the dominant K-feldspar is orthoclase microperthite. The composition of plagioclase is about An_{12} and its alteration to a fine-grained white mica is typical.

The greisenised rocks are rather heterogeneous in texture and composition but essentially consist of quartz, sericite, muscovite, topaz, tourmaline, apatite, lithium mica and cassiterite. The quartz shows evidence of recrystallisation and metasomatism. Two phase liquid-gas fluid inclusions are ubiquitous in these metasomatic quartzes. The average modal compositions of the granite and greisen are presented in table 1.

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TABLE 1

Modal analyses of granites and greisens

	Unaltered granite	Greisen
Quartz	31.5	40.65
Potash feldspar	35.18	
Plagioclase	21.69	
Mica	7.55	42.73
Tourmaline	3.39	1.80
Topaz	0.50	14.00
Apatite	0.13	0.74
Cassiterite		0.05
Opaques		0.03

The modal compositions given above are the average of modal analyses of four specimens of each rock type.

3. Chemical composition of the rocks

The chemical composition and normative constituents of nine granites and four greisens are presented in tables 2 and 3. As far as the major element composition is concerned, all the granite varieties are very similar. The average St. Michael's Mount granite shows no unusual compositional features from other Cornubian granites (Exley and Stone 1964) except for the abundance of B_20_3 and F. The average boron content of all granites is estimated by Turekian and Wedepohl (1961) to be approximately ten parts per million, however the average apparently unaltered granite of St. Michael's Mount contains 1460 parts per million which is more than one hundred and forty six times greater. This abnormal concentration of boron was probably produced by partial melting of sediments with high boron content (Hall 1971). The normative compositions of the granites are all very similar, although because of the high boron and fluorine concentrations, a certain amount of error is to be expected in the calculation of these norms. When plotted on a normative quartz-orthoclase-albite triangular diagram (fig 1) the granites fall near the ternary minimum for a water pressure of approximately 1000 bars (Tuttle and Bowen 1958), plotting very close to the other granites of the region.

To investigate the gradational nature of the granite-greisen transition the same experimental method as that of A. Hall (1971) was used, i.e. the chemical variation across a cross-section of the rock adjacent to a mineralised quartz vein was measured. The sample used for this purpose was a large piece of rock (35 cm x 17 cm) containing a mineralized quartz vein with coarse mica selvedge in contact with

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TABLE 2

Chemical analyses of St. Michael's Mount granite

	SM 46	SM 48	SM 50	SM 52	SM 54	SM 56	SM 58	SM 60	SM 62
SiO_2	73.30	72.62	72.93	74.18	73.39	73.48	73.01	73.99	72.81
Al_2O_3	15.02	15.49	15.11	14.54	14.66	15.12	15.50	14.60	15.34
Fe_2O_3	0.24	0.21	0.22	0.26	0.26	0.25	0.10	0.24	0.38
FeO	0.77	0.69	0.74	0.79	0.76	0.76	0.78	0.90	0.79
MgO	0.13	0.14	0.10	0.15	0.15	0.10	0.10	0.15	0.14
CaO	0.40	0.35	0.35	0.44	0.38	0.38	0.50	0.39	0.42
Na_2O	3.24	3.26	3.46	3.18	3.07	3.37	3.79	3.00	3.30
K_2O	4.53	4.80	4.61	4.35	4.50	4.19	4.46	4.90	4.46
H_2O+	0.64	0.64	0.60	0.57	0.65	0.76	0.67	0.61	0.64
H_2O-	0.17	0.20	0.17	0.10	0.15	0.22	0.16	0.19	0.14
TiO_2	0.11	0.07	0.10	0.11	0.11	0.07	0.09	0.21	0.09
P_2O_5	0.27	0.28	0.25	0.23	0.29	0.28	0.23	0.25	0.31
MnO	0.05	0.03	0.05	0.04	0.04	0.06	0.04	0.04	0.07
Li_2O	0.11	0.10	0.10	0.11	0.09	0.12	0.09	0.10	0.13
B_2O_3	0.47	0.48	0.40	0.46	0.68	0.59	0.46	0.20	0.51
F	0.52	0.49	0.69	0.46	0.60	0.49	0.47	0.48	0.47
-O=F	0.22	0.21	0.29	0.19	0.25	0.21	0.20	0.20	0.20
Total	99.75	99.64	99.59	99.78	99.53	100.05	99.87	99.89	99.80

Norms

	SM 46	SM 48	SM 50	SM 52	SM 54	SM 56	SM 58	SM 60	SM 62
Qtz.	36.33	34.66	34.48	38.02	37.62	37.13	35.57	37.04	35.82
Or.	26.76	28.36	27.24	25.70	26.59	24.76	26.35	28.95	26.35
Ab.	27.40	27.57	29.26	26.90	25.96	28.50	27.83	25.37	27.91
An.	0.22	0.09	0.10	0.68	0.01	0.06	0.98	0.30	0.06
Fs	1.13	1.03	1.10	1.13	1.07	1.22	1.50	0.89	1.12
En	0.32	0.35	0.25	0.37	0.37	0.25	0.25	0.37	0.35
Mt.	0.35	0.30	0.32	0.38	0.38	0.36	0.14	0.35	0.35
Il.	0.21	0.13	0.19	0.21	0.21	0.13	0.17	0.40	0.17
Co.	4.71	4.97	4.39	4.35	4.54	5.02	4.90	4.25	5.06

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TABLE 3

Chemical analyses of St. Michael's Mount greisens

	SM64	SM66	SM68	SM70
SiO_2	71.85	72.78	72.34	72.24
$A1_2O_3$	14.89	14.04	14.69	15.01
Fe_2O_3	0.36	0.24	0.23	0.32
FeO	2.98	3.35	3.14	3.19
MgO	0.16	0.10	0.08	0.15
CaO	0.66	0.57	0.50	0.57
Na_2O	0.13	0.13	0.13	0.14
K_2O	4.39	3.68	3.62	3.23
H_2O+	1.01	1.10	0.86	0.97
H_2O-	0.17	0.20	0.24	0.11
TiO_2	0.19	0.07	0.11	0.15
P_2O_5	0.48	0.41	0.35	0.49
MnO	0.37	0.33	0.39	0.34
LiO_2	0.64	0.74	0.89	0.72
B_2O_3	0.22	0.41	0.29	0.04
F	2.18	2.54	3.15	3.60
-0=F	0.87	1.07	1.32	1.51
Total	99.76	99.62	99.69	99.76

greisen, the latter passing into a feebly porphyritic tourmaline-rich granite, which in its turn grades into unaltered granite porphyry. The rock was cut into six slices parallel to the contact between the vein and the greisen representing the greisen and granite portions. These slices were crushed and analysed by wet chemistry to give a continuous profile of chemical variation. The results are shown in Fig 2.

4. Conclusions

This preliminary study of the St. Michael's Mount granite and associated greisens has revealed that at least four different granitic facies are present in the cupola. However, before coming to any conclusion about the genetic and textural relationships between these facies, detailed mapping and careful sampling of each for analysis is essential. What can be concluded is that in comparison with other Cornubian granites the average St. Michael's Mount granite is the richest by far in tourmaline (and hence boron).

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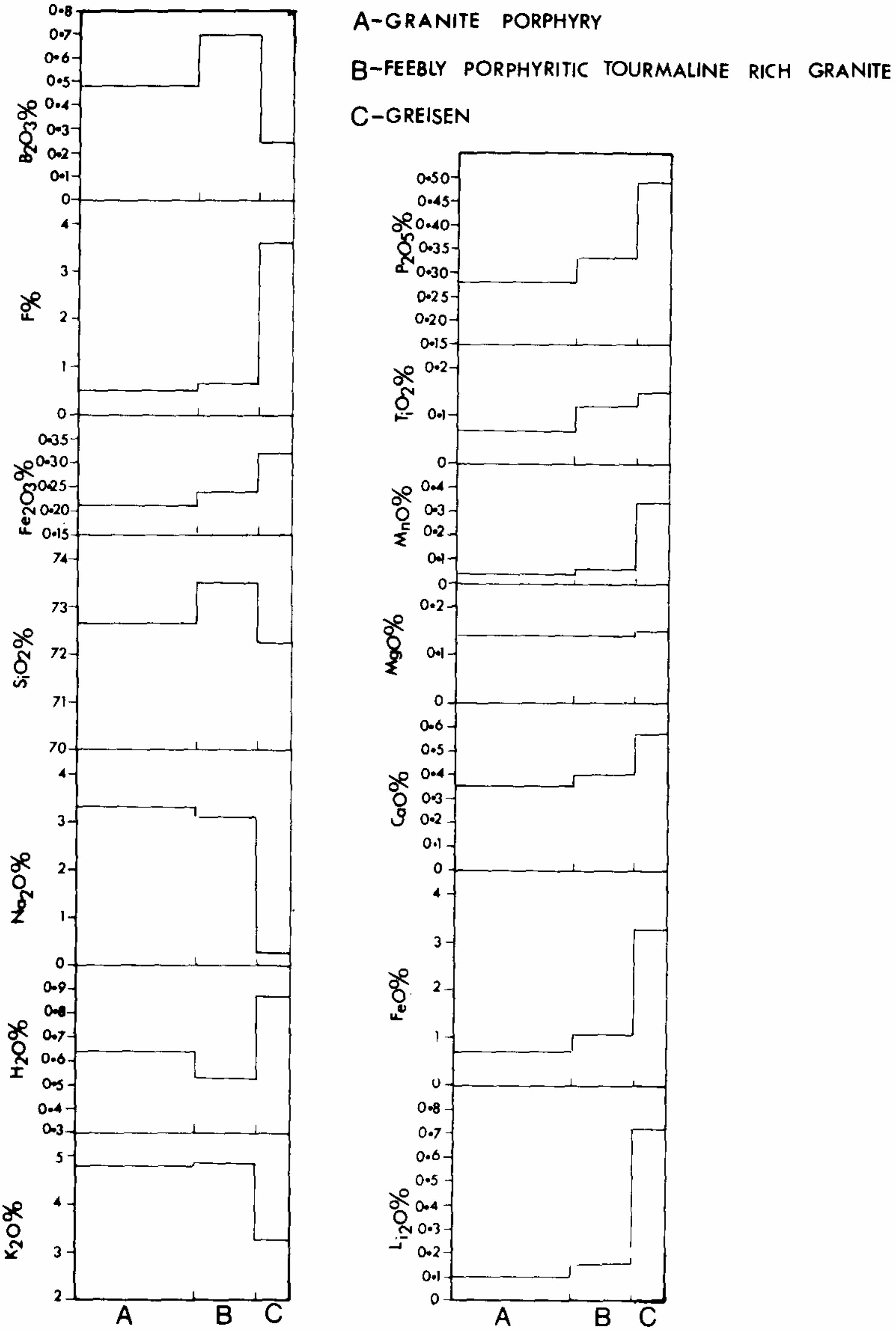
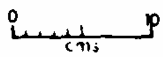


Fig 2.

The variation in chemical composition across a greisenised granite adjacent to a mineralised quartz vein.

SCALE:-



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The St. Michael's Mount greisens are unique; firstly, they contain over thirteen percent modal topaz, indicating an abnormal concentration of fluorine in the greisening solutions; secondly, the tourmaline content is less than that of the host granite. An explanation could be that some tourmalines lost their stability during the process of greisenisation and Were carried away by the solutions. Further characteristics are a fairly high lithium content (due to a high lithium mica content) and a somewhat high phosphorus content.

Even though the above characteristics are quite striking, it must be remembered that all greisens show widely varying compositions (except for the general vigorous depletion of sodium). This fact can only be explained in terms of the chemistry of each original altering solution and the differing physio-chemical conditions of each environment.

ACKNOWLEDGEMENT. The author would like to thank Professor R.A. Howie for critically reading the manuscript.

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NOTES ON MARBLE AND CALC-SILICATE ROCKS FROM DUCHY PERU BOREHOLE, NEAR PERRANPORTH, CORNWALL

by A.J.J. Goode and R.J. Merriman

1. Introduction

Calc-silicates have been described by Collins (!874) from Great Retallack Mine, by Warington Smyth (1882) from Duchy Peru Mine, and have been further documented by Flett in the same area (in Reid and Scrivenor 1906). These early accounts were reviewed by Henley (1971), who described the occurrence of hedenbergite and sphalerite associated with the Perran Iron Lode.

In 1973-4, an inclined borehole, drilled approximately 400 m west of Duchy Peru Mine to examine the Perran Iron Lode and associated geophysical anomalies, intersected 62.50 m of altered calcareous rock at 141.73 m including 39.77 m of white marble. The borehole is sited in the Meadfoot Beds close to the boundary with the Gramscatho Beds, and the lime-rich rocks occur in a sequence of grey chloritic slates cut by sporadic elvan (quartz-porphyry) dykes.

2. Marble

The marble is fine-grained and white in colour, with pale grey banding at 65^0-80^0 to core length apparently of sedimentary origin, it is typically saccharoidal and consists of anhedral granoblastic calcite averaging O. 15 mm across, containing scattered areas of granoblastic quartz which may be linked by trails of sub-idioblastic quartz grains (IGS Slide No. E 49691). Flakes of biotite are found within the marble adjacent to argillaceous intercalations. The latter consist of lamellae of finely intergrown chlorite, muscovite, quartz and feldspar with discontinuous yellow-brown lamellae rich in finely granular masses of rutile (E49690). The marble is a hard cohesive rock and lengths of core in excess of I m were recovered, suggesting a lack of jointing. Thinly-bedded argillaceous layers occur between 192.48 m and 193.09 m and irregular segregations of sphalerite between 194.46 m and 194.61 m. Samples of marble dissolved in dilute acetic

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acid failed to yield any microfossils and it is possible that the local conditions of deformation and metasomatism removed all traces of any existing fauna.

3. Calc-silicate rocks

The marble appears to be closely associated with an envelope of calc-silicate-bearing rocks (known locally as calc flintas). They occur above the marble at 141.73 m and 154.84 m and below it at 194.61 m and 197.97 m. Thin veins and lenticular segregations of calc-silicates are present in the underlying grey slates to 204.22 m. Similar rocks from St. Pirans Church (SW. 7715. 5645) and Great Retallack Mine (SW 790.560) were described by Flett as greenstones (in Reid and Scrivenor 1906).

The calc-silicate rocks are made up largely of a green pyroxene, which X-ray powder patterns suggest is a hedenbergitic variety (Zwaan 1954). A finely granular rock at 195.38 m is composed of colourless decussate epidote interbanded with fibrous radiating hedenbergite. Dark green blades of hedenbergite up to 30 mm long occur at 195,91 m, forming radiating sheaves locally intergrown with granular aggregates of red-brown grossular and patchy calcite. Xenoblastic aggregates of grossular also occur in medium-grained, pale. green rock at 196.29 m. In thin section (E 49693) the garnet is pale pink and usually birefringent, in both respects resembling grossular in garnet rock (IGS Museum Collection No, MR 1891; E3665)described by Flett from St. Pirans Church. Yellow-brown andradite garnet, occurring as well-formed crystals up to 15 mm in diameter, has developed with sphalerite segregations at the marble-calc-silicate junction. Axinite is widely distributed and above the marble occurs as pale purple-grey platy crystals up to 10 mm across. Minor pyrite and silver grey marcasite are associated with the sphalerite-andradite intergrowth. An oblique shear cuts the core below the marble-calc-silicates junction and is coated with a blue-green asbestiform amphibole which X-ray diffraction suggests is probably an actinolitic variety.

The grey laminated slate below the calc-silicate rock has been invaded for approximately 6 m by thin, pale green veins and lenses composed chiefly of hedenbergite. The sediments, originally perhaps calcareous or dolomitic shales, have been altered to hornfelsed slates consisting of bands of granoblastic andesine alternating with thin lamellae rich in granular diopside, epidote and sphene (E 49694).

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Portable X-ray fluorescence equipment has revealed the presence of minor tin mineralisation throughout the calc-silicate rock. It appears to be disseminated and may prove to be similar to the finely divided cassiterite in the calc-silicate rocks of the Mulberry area, Lanivet (Hosking 1969). Alternatively the tin may have entered silicates such as grossular or sphene under metasomatic conditions (Sharwaki and Dearman 1966). Cassiterite has also been recorded within the Perran Iron Lode by Sabine (1968).

4. Discussion

The marble may be correlated with lime-rich horizons in the Meadfoot Beds on the N limb of the Newquay anticline, which gave rise to the calc-silicate rocks of the St. Columb-Withiel district. The only limestones with a comparable thickness found locally have been described by Sadler (1973) in the Veryan area.

It is interesting to note that in 1903 J.B. Scrivenor recorded in the Gramscatho Beds "Fragments of Plymouth Limestone abundant" from ½ km S of Ennis (SW 835.522) to 1/2 km N of Killiserth (SW 853.525) and in the Meadfoot Beds 1 km NE of New Mills (SW 900.532) "Mid Devonian Limestone". These localities are close to the strike of the Duchy Peru marble projected in an ESE direction.

The calc-silicate rock which crops out just west of St. Pirans Church (shown on Geological Survey 1:50 000 map as greenstone) bears a close resemblance to that seen in the Duchy Peru core. The marble, being closely associated with the calc-silicate rocks in the borehole might also crop out nearby where it is not concealed by Blown Sand.

The biotite in the marble and the thermal spotting found in the vicinity of St. Pirans Church suggests the existence of a north-easterly subsurface extension of the Cligga Granite.

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Upper Devonian purple and green slates in north Cornwall (Abstract):

by Anthony P.Beese

A study of basinal sediments of the Polzeath Slates in the Camel Estuary demonstrates that a lithological correlation is possible despite the intense deformation. The structure of the slates is a synclinorium with overturned folds developed on the north and south limbs. There is no major faulting.

A maximum thickness of 250 m of slates can be subdivided into three divisions. The lowest division consists of 55m of purple Slates with common siltstone bands. The middle division comprises 30 m of purple slates with two l.5m thick green slate bands containing dark grey slate bands; and 80m of purple slates with common 30-80 cm green slate bands which cannot be correlated precisely. The upper division consists of 80-90 m of predominantly purple slates with occasional chert, ash, green slate and siltstone hands. The middle and upper divisions contain ostracods and a goniatite indicating a low and middle Famennian age, the underlying Frasnian green Harbour Cove Slates contain conglomerates and tuffs to the north, and black goniatite slate bands to the south.

A possible interpretation for the deposition of the Polzeath Slates suggests a subsiding and uneven sea floor with a volcanic rise to the north producing the volcanic and turbiditic element in an otherwise pelagic succession.

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Conodonts from Middle/Upper Devonian boundary beds at Barton Quarry, Torquay (Abstract):
by Christine Castle

During a study of Devonian conodonts from the Torquay area, faunas have been found in Barton Quarry (SX 91256710), now the Torquay Holiday Village, that indicate an age about the Middle-Upper Devonian boundary. The limestone of the quarry is massive, bioclastic and crystalline, and contains corals and stromatoporoids. This is the type locality for the coral *Phillipsastrea hennahi*; the goniatite *Wedekindella brilonense* has also been recorded from here. An upper Givetian age was suggested on this macrofaunal evidence.

The conodont zones concur in part with this age. The upper Givetian *varcus* Zone is represented by *Polygnathus varcus*, *P. xylus* and *P. tirnorensis* with *P. linguiformis*. The uppermost *varcus* Zone is suggested by samples with *P. decorosus* transitional to the genus *Schmidtognathus,* with *P. ovatinodosus, P. xylus* and *Icriodus eslaensis*. The upper part of the S. *hermanni - P. cristatus* Zone (a critical zone for the determination of the Middle - Upper Devonian boundary) is indicated by faunas of *S.* cf. *peracutus*, *S.* cf. *pietzneri* and occasional *P. cristatus*, with *P. dubius, P. ovatinodosus* and *P. xylus*. The lowermost part of the Frasnian *asymmetricus* Zone is represented by a sample with *S.* cf. *hermanni, S.* cf. *pietzneri, P. dubius, P. ovatinodosus, P. xylus* and *P. cristatus* transitional to *P. asymmetricus.* Thus, on the conodont evidence, the limestone of Barton Quarry appears to have an age ranging from the Givetian into the Frasnian.

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ADDITIONAL CONODONTS FROM NEAR THE MIDDLE/ UPPER DEVONIAN BOUNDARY IN NORTH CORNWALL A PROGRESS REPORT

by N.A. Mouravieff

Abstract. The section at Marble Cliff, Trevone, studied by Kirchgasser has been resampled and the observations extended to older and younger strata. The recognised conodont zones range from the *varcus* Zone 1971 to the lower *asymmetricus* Zone. The conodont fauna is abundant in the limestone beds of turbiditic origin and is composed of both shallower-water and deeper-water forms. The section reveals a considerable thickness of each of the conodont zones and is valuable for a more detailed observation of the evolution of the conodonts in this interval near the Middle/Upper Devonian boundary.

The rock succession in the Trevone area has been reviewed by Gauss & House (1972) for the stratigraphy. Tucker (1969) has described the sedimentology of the Marble Cliff Beds and Kirchgasser (1970) made the first conodont study.

Seventy-four samples (1-2 kg each) were processed and produced more than ten thousand complete and broken conodonts. Four conodont zones were identified.

1. *varcus* Zone (1971). (Thickness: uncertain; 1 sample; 216 conodonts).

Kirchgasser (1970) has already recognised the *varcus* Zone (1962) at Trevone Bay. It appears that his bed No. 31 contains a fauna of the *varcus* Zone 1971 identified by the abundance of a conodont of the varcus group, that is, *Polygnathus rhenanus.*

2. Problematical interval of the *hermanni-cristatus* Zone (Thickness: 12m; 20 samples; 1352 conodonts).

This interval starts above the dolerite sill which is observed at the beginning of the Marble Cliff section, in the first layers of this interval *Schmidtognathus wittekindti* is present. According to Ziegler (1966, 1971) this species appears in the Upper *hermanni-cristatus* Zone in association with *Polygnathus cristatus*. But this latter species appears only higher in the Marble Cliff section.

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3. Upper *hermanni-cristatus* Zone. (Thickness: 11 m; 14 samples; 1947 conodonts).

Recognised by *Polygnathus cristatus* present in practically every sample of the zone, associated in some cases with *Schmidtognathus hermanni*.

4. The Lowermost *asymmetricus* Zone. (Thickness: 40 m; 32 samples; 4290 conodonts).

Starts five metres below bed No. 1 of Kirchgasser includes his bed No. 12 and extends at least ten metres above his bed No. 12. The zone is identified by the occurrence of *Polygnathus asymmetricus*. *Polygnathus cristatus* is still regularly encountered throughout the whole zone. New forms appear, such as *Polygnathus* new sp. which is distinguished from the other species of the genus by the asymmetry of its basal cavity being entirely located in the posterior half of the flat platform. A more or less typical specimen of this new species is figured by Kirchgasser (1970, pl. 63, fig. 8ab; *not Palmatolepis transitans*).

A new species of the genus *Palmatolepis* is characterised by an asymmetrical basal cavity of the same type as that shown by *Palmatolepis? disparalvea* but it differs from this latter species in having a lobe which is not well differentiated and by having a well differentiated central node.

Frequent in this zone and extending higher in the next zone is *Polygnathus dengleri* which shows variations which could lead to subspecific discriminations.

In the upper eight metres of the zone *Polygnathus pennatus* (see Huddle 1970) is abundant and regularly encountered; so this species could be used to distinguish an upper part of the zone. In the same upper eight metres *Polygnathus norrisi* occurs in association with a new species of *Ozarkodina* which has the same type of ornamentation as the platform of *P. norrisi*. Uyeno thinks that this new *Ozarkodina* could be the ozarkodinian element of the *P. norrisi* apparatus (personal communication).

5. The Lower *asymmetricus* Zone. (Marble Cliff; Porthmissen Bridge plus Longcarrow Cove sections: 18+65 m. thick; 7+8 samples; 1173+1386 conodonts).

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Includes beds Nos. 13-20 of Kirchgasser and extends through Porthmissen Bridge towards Longcarrow Cove.

The contact between the preceding Lowermost *asymmetricus* Zone and the Lower *asymmetricus* Zone seems to be a fault-contact and must be re-examined in detail. The lowest bed actually known of the zone is bed No. 13 of Kirchgasser which contains *Ancyrodella rotundiloba rotundiloba.*

Conclusions: The comparison with the sections of the Rhenish Slates Mountains on which Ziegler (1966) based his conodont zonation for the considered interval shows that the thickness of each of the recognised conodont zones is 13 to 50 times thicker in Cornwall. Thus the conditions are better for detailed observation of conodont evolution and for distinguishing evolutionary steps and phyletic relationships.

There is no doubt that the Cornish conodonts recovered from turbidites have been, at least partially, laterally displaced but there is no evidence of vertical reworking which could have mixed older and younger forms.

Finally, the fauna includes shallower-water and deeper-water forms, thus giving a more complete range of forms than does the fauna of the shallower-water deposits of the "Assise de Fromelennes" in the Ardennes.

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CARBONIFEROUS SUCCESSIONS IN GERMANY AND IN SOUTHWEST ENGLAND

by S.C. Matthews

Abstract. The chief characteristics of the Carboniferous succession in the Rheinisches Schiefergebirge and Harz regions of Germany are briefly described. The Carboniferous in Devon and Cornwall is comparable in numerous respects. But there is one major difference to be discussed: nothing in the German segment of the Variscan foldbelt compares with the largely non-marine Bude Formation, of early Westphalian age. It is argued that this late influx of sediment found accommodation in a sink developing adjacent to the growing granite batholith. If this is the case, it follows that emplacement of the batholith was a process which took place over a period of time (early Westphalian to end-Carboniferous approximately) of the order of 30 million years. Some of the implications of this proposal are examined.

1. The Carboniferous in Germany

A brief sketch map must suffice here. Fuller information (in English) is available in Meischner (1971) and Paproth (1971). It will be noted that this present account draws heavily on Meischner's paper.

The Carboniferous of the Rheinisches Schiefergebirge and the Harz is basinal, thin and for the most part "starved". The Lower Carboniferous is divided into three stages or Stufen, successively the *Gattendorfia*-Stufe (cu I), the *Arnmonellipsites*-Stufe (cu II) and the *Goniatites*-Stufe (cu III). Originally, "cu" was a map-symbol, just as "d2" is a map-symbol in Great Britain. The *Gattendorfia*-Stufe is very thin. At its type expression, in the Hangenbergkalk, it comprises two metres of nodular limestone. The figures given by Paproth (1971) would suggest that the total thickness of the starved succession (cu I, cu II plus a low part of cu III) would rarely exceed one hundred metres. In some places it can be very much less. Meischner and Schneider (1970) have described a section in which much of the Upper Devonian and a fair part of the Lower Carboniferous are represented by a one metre thickness of rock.

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Coarse, feldspar-rich clastics occur early in the Lower Carboniferous in areas close to the southeastern margin of the Rheinisches Schiefergebirge and also in the southwestern part of the Harz - the Tanner Greywacke is one example there. These sediments were derived from sources in crystalline rocks to the south and were trapped in an area south of what Meischner (1968,1971) has called the "mid-geosynclinal rise". The effects of this rise can be traced over a distance of 250 kin, from the Dill Syncline east-northeastward through the Kellerwald and into the Harz. Successions in both Devonian and Carboniferous are especially thin on the rise, yet they can enclose occasional developments with a benthonic fauna (e.g.' Kellerwald-Quarzit). Basic volcanics (these including the Deckdiabas in cu II), common along the general strike of this major, narrow tectonic feature, are much less in evidence elsewhere.

Northwestward of the rise, the starved basin sequence includes bedded chert, which occupies much the same range of age as the Deckdiabas volcanism. Higher in the Lower Carboniferous, calcareous turbidites enter the basinal succession. At least four separate streams can be distinguished. Two (the Rhenaer-Kalk and the Posidonienkalk: Meischner 1962) came from the east or southeast, another (the Hellefelderkalk: see references in Meischner 1971) is suggested to have been derived from a source (Devonian massive limestone) within the Carboniferous basin, and the fourth (Kulm-Plattenkalk: Meischner 1971, see especially fig. 7) is the only one to have come from the region of the northern shelf. This last is a rare example of transport of material from the northern shelf into the basin. Franke, Eder and Engel (1975) have made a sedimentological analysis of the carbonates deposited near the margin of the northern shelf. This early Carboniferous margin lies well to the northwest of the shelf margin of mid Devonian time - a shift of some 50 km across the strike is involved.

The major influx of greywacke (Kulmgrauwacke) came from the southeast, spilled over the mid-geosynclinal rise and spread out in the previously starved basin. These sediments are more mature than the feldspathic sands which were trapped in the southeast during earlier Carboniferous time. The thickness of the Kulmgrauwacke can range up to 1000 m. The base of the sequence is progressively younger northward and near the northern limit of the Rheinisches Schiefergebirge is finally of early Namurian age. Farther to the north, there is the thick Namurian-Westphalian succession of the Ruhr coalfield; but within the foldbelt, the youngest-surviving rocks are early Namurian greywackes.

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2. Comparable developments in the Carboniferous of southwest England

In Devon and Cornwall, where the Carboniferous rocks are much more drastically deformed, stratigraphical and sedimentological information has accumulated relatively slowly. Nevertheless, it has become possible to see that the succession resembles the German one in a number of ways. The Lower Carboniferous rocks are generally fine-grained and of basinal character. It is difficult to make estimates of thickness, but reliable information from a few localities - e.g. Chudleigh, where House and Butcher (1973) have shown that the thickness of the Lower Carboniferous is less than 40 m - encourages the belief that the succession is a thin one, and much thinner than the sequence of the same age on the shelf to the north, Feldspathic sands are known to occur at southern localities, e.g. St. Mellion (Matthews 1966: note there a mention of possible resemblance to German records of material derived from the Mitteldeutsche Schwelle) and Chudleigh, where the first elastics in the succession are feldspathic (House and Butcher 1973). Basic volcanics - at Tintagel (Freshney, McKeown and Williams 1972) and in the Teign Valley (Chesher 1968) for example - are of much the same age as the Deckdiabas. In southwest England as in Germany there is a widespread development of chert in cu II. The fine-grained succession above the chert development includes thin limestones in many places, but it is not yet possible to suggest regional patterns of distribution of all of these carbonates. Matthews and Thomas (1974) have suggested that the Westleigh Limestone is a fan of calcareous turbidites, whose material includes debris derived from the shelf.

The thin basinal succession is followed, as in Germany, by a major sequence of lithic greywackes. Crackington Formation greywackes first arrived, according to our present information, rather later than earliest Kulmgrauwacke - in early Namurian rather than late Visean. It is known that the Crackington Formation was deposited from currents flowing east-west (Freshney and Taylor 1972), but there is, as yet, no clear indication of the location of the source from which the sediment came.

Several points of the evidence suggest that through Dinantian time and into early Namurian time, the Carboniferous successions of these two regions were developing along near-parallel courses. Later, however, a major difference emerged: in the German segment of the

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Variscan foldbelt there is nothing which can be compared with the early Westphalian Bude Formation.

3. The Bude Formation and the granite batholith

The Bude Formation (its general pattern of outcrop is shown in Edmonds, McKeown and Williams 1975, fig. 1) is a succession of massive sandstones, slumped beds and shales (King 1966, Freshney and Taylor 1972, Freshney, McKeown and Williams 1972). Freshney and Taylor (1972, pl.2) have recorded a thickness of 1300 m. It is now widely accepted that much of the succession is non-marine (Goldring and .Seilacher 1971; Freshney and Taylor 1972). There need be no suggestion that the Bude Formation is in any way to be associated with the coarse feldspathic sandstones found at lower Carboniferous horizons in the southern part of the region: the Bude Formation sediments were delivered from a sector between northwest and northeast (Freshney and Taylor 1972). The evidence from this highest surviving part of the stratigraphic sequence indicates that the earlier pattern of palaeogeography, broadly consistent with that in Germany, had been upset. The basin in which the Crackington Formation had been deposited from axial flows was now choked, and possibly to some extent uplifted and restricted (as the suggestion of non-marine conditions would imply). And yet sediment of this character accumulated to considerable - by southwest England standards - thickness and survived within the foldbelt. In Westphalian A, it appears, there was a radically new influence at work.

Burne (1973) has proposed that the major event reflected in these changes was a collision of continents, which closed what had previously been a small oceanic basin. The case for continental collision remains unproven, however. So too does the case for oceanic affinity of anything in the Upper Palaeozoic geology of southwest England. Other possible explanations of the presence of the Bude Formation deserve examination, and it would be reasonable to prefer an explanation which applies to southwest England alone, rather than to the whole Variscan tectogene.

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It is conceivable that the new influence was the first growth of the granite batholith, inducing development of a marginal sink and thus providing accommodation for the influx of sediment from the north. The nature of marginal sink structures has been described, on the basis of experimental models, by Ramberg (1967), Bridgwater Sutton and Watterson (1974) have discussed relationships of this kind, as developed at deeper crustal levels. Sannemann (1963,1968), following Trusheim (1960), has dealt with the analogous case of salt domes, which build into their cover first a primary peripheral sink, at salt-pillow stage, then later a secondary peripheral sink, at diapir stage. The Bude Formation may be thought to have been deposited in a depression related to the early development of the batholith. Later, it was folded (along with the other Carboniferous rocks of north Cornwall) and then cut by low angle normal faults which were active during emplacement of the high level Bodmin granite (Freshney 1965; Freshney, McKeown and Williams 1972).

Some of the implications of such a proposal should be examined:

1. If the presence of the batholith is first in evidence in Westphalian A (i.e. approximately 310 m.y. according to Francis and Woodland 1964) and if the last crystallization, more or less consistently indicated by radiometric dating of samples from the several granite cupolas, is at 280 m.y. approximately, then it follows that the batholith was in process of emplacement through a period of time of the order of 30 million years. One might wish to compare the rather higher figures (50-70 million years) recently proposed for the Coastal Batholith of Peru (Pitcher 1975).

2. If the suggestion that a major granite pluton was already gathering during Westphalian A is sound, it is implied that the batholith was already in play during the time when the south-facing fold-structures (involving Bude Formation) seen on the north Cornwall coast were produced.

3. Do the granites themselves supply any information to support these proposals? Probably they cannot: one learns from numerous publications (e.g. Exley and Stone 1966; Stone 1971, 1975; Floyd 1972) that the granites, at levels currently exposed, are highly-evolved bodies in which any indication of earlier stages of crystallization is unlikely to have survived. Evidence bearing on the question of the beginnings of emplacement must be sought in effects built into the cover.

4. It will be noted that this suggestion of a stratigraphic effect related to granite replacement is in some respects reminiscent of Krebs and Wachendorf's (1973, 1974) proposals on what they take to have been the controlling influences on the development of the Central European Variscides. It should be emphasized, however, that in southwest England, such an effect becomes evident only at a relatively late stage in the tectonic history. The earlier development of the Variscan geology of Devon and Cornwall, in which there is strong resemblance to what is seen in the Rheinisches Schiefergebirge and the Harz, is better discussed in terms of control by major fractures (Matthews 1977).

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British Triassic Palaeontology: a supplement

by G. Warrington

Since the completion of the manuscript of the writer's paper on British Triassic palaeontology (*Proc. Ussher Soc.*, **3**, 341-353; 1976) the following works dealing with or including aspects of the palaeontology of the British Trias have appeared:

IRELAND, R.J., POLLARD, J.E., STEEL, R.J. and THOMPSON, D.B. 1978. Intertidal sediments and trace fossils from the Waterstones (Scythian - Anisian?) at Daresbury, Cheshire. *Proc. Yorks. geol. Soc.*, **41**, 399-436.

POLLARD, J.E. 1976. A problematic trace fossil from the Tor Bay Breccias of south Devon, (written discussion to paper taken as read). *Proc. Geol. Ass.*, **87**, 105-108.

POLLARD, J.E. 1976. A problematic trace fossil from the Tor Bay Breccias of south Devon, (written discussion to paper taken as read). *Proc. Geol. Ass.*, **87**, 105-108.

RIDGWAY, J.M. 1976. (reply by the author to J.E. Pollard). *Proc. Geol. Ass.*, **87**, 108-109.

NICKLESS, E.F.P., BOOTH, S.J. and MOSLEY, P.N. 1976. The celestite resources of the area north-east of Bristol with notes on occurrences north and south of the Mendip Hills and in the Vale of Glamorgan: Description of 1:25,000 resource sheet ST 68 and parts of ST59.69, 79.58, 78.67 and 77. *Miner. Assess. Rep. Inst. Geol. Sci., No*. **25,** 83pp.

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PALYNOLOGICAL EXAMINATION OF TRIASSIC (KEUPER MARL AND RHAETIC) DEPOSITS NORTH-EAST AND EAST OF BRISTOL

by G. Warrington

Abstract. Mudstone samples from the Keuper Marl in the Cromhall - Yate - Bitton area north-east and east of Bristol are virtually devoid of palynomorphs. The Rhaetic Westbury and Cotham beds near Yate have yielded palynomorph assemblages comprising miospores and organic-walled microplankton, comparable with those recorded from the same facies elsewhere in southern England.

1. Introduction

In connection with the assessment of celestite resources to the north-east and east of Bristol, Avon (Nickless, Booth and Mosley 1976), 84 samples were processed and examined for palynomorphs. The majority (78) of the samples originated from the Keuper Marl (including the Tea Green Marl; hereafter, TGM) and were examined to determine whether there was palynological evidence for the correlation of the main celestite-bearing unit (the Severnside Evaporite Bed; hereafter, SEB) in the Keuper Marl of this district. Preparations from this material were, however, devoid of stratigraphically significant palynomorphs and, in most instances, were totally barren. The six remaining samples, four of which yielded palynomorph assemblages, originated from the Rhaetic Westbury and Cotham beds.

2. Samples from the Rhaetic

Cotham and Westbury beds samples were examined from the following units and depths in two boreholes sited at Wapley and Pucklechurch (see Nickless, Booth and Mosley, 1976, for borehole logs and locations; page numbers cited below in parentheses after the borehole numbers refer to that account).

Borehole ST 78 SW 12, Cliff Farm, Wapley (p. 62)
- Cotham Beds: 2.70 m, 3.30 m, 3.85 m
- Westbury Beds: 4.00 m, 6.20 m

Borehole ST 67 NE 59, Park Farm, Pucklechurch (p. 52)
- Westbury Beds: 8.00 m.

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Depth in borehole (metres)	2.70	3.30	3.85	4.00
Sample number: SAL	2501	2482	2483	2484
MIOSPORES:				
Triancoraesporiles reticulatus Schulz 1962	+			
Deltoidospora neddeni (Potoni6) Orbell 1973	+			
Convolutispora microrugulata Schulz 1967	?			
Cornutisporites cf. seebergensis Schulz 1962		+		
Tsugaepollenites? pseudomassulae (Madler) Morbey 1975	*	+	*	
Kraeuselisporites reissingeri (Harris) Morbey 1975	+	+	+	
Lunatisporites rhaeticus (Schulz) Warrington 1974		+	*	
Vesicaspora fuscus (Pautsch) Morbey 1975	+	+	+	+
Ricciisporites tuberculatus Lundblad 1954	7.0	7.5	4.0	14.0
Rhaetipollis germanicus Schulz 1967	3.5	3.0	4.5	7.5
Ovalipollis pseudoalatus (Thiergart) Schuurman 1976	4.0	6.0	7.0	3.0
Classopollis torosus (Reissinger) Balme 1957	38.0	49.5	33.5	29.5
Alisporites thomasii (Couper) Nilsson 1958	3.5	1.5	*	+
Zebrasporites laevigatus (Schulz) Schulz 1967	?	+		+
Deltoidospora hallii Miner 1935	+		+	+
Perinosporites thuringiacus Schulz 1962	?			+
Cycadopites sp.	+			+
Alisporites sp.		3.0	+	2.5
Microreticulatisporites fuscus (Nilsson) Morbey 1975		+		+
Limbosporites lundbladii Nilsson 1958		?		+
Acanthotriletes varius Nilsson 1958		+		+
Quadraeculina anellaeformis Maljavkina 1949			+	
Leptolepidites argenteaeformis (Bolkhovitina) Morbey 1975			+	
Carnisporites anteriscus Morbey 1975			+	
Acanthotriletes ovalis Nilsson 1958			+	
Lycopodiacidites sp.				+
ORGANIC-WALLED MICROPLANKTON:				
Micrhystridium sp.	4.0	4.0	+	1.5
Rhaetogonyaulax rhaetica (Sarjeant) Loeblich & Loeblich emend. Harland, Morbey & Sarjeant 1975	34.5	21.0	42.5	35.0
Tasmanites cf. suevicus (Eisenack) Wall 1965			+	
Cymatiosphaera polypartira Morbey 1975			+	*
`Hystrichosphaeridium' langi Wall 1965 sensu Harland 1975				*

Table 1: Composition of palynomorph assemblages from the Rhaetic, Borehole ST 78 SW 12, Cliff Farm, Wapley

(Relative abundances expressed as percentages based upon counts of 200 specimens; + = 0.5%, * = 1 %

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The lowest Westbury Beds sample from the Cliff Farm Borehole, Wapley, and that from the same unit in the Park Farm Borehole, Pucklechurch, were barren. The compositions of the palynomorph assemblages recovered from the remaining Cliff Farm Borehole samples are given in Table 1. The preparations are registered and stored in the SAL collection at the Institute of Geological Sciences, Leeds.

In composition and general character the assemblages are comparable with those documented by Orbell (1973) from the Westbury Beds and lower part of the Cotham Beds in the Upton Borehole, Oxfordshire, and by the writer from the same units in the Steeple Aston and Withycombe Farm boreholes, Oxfordshire (Warrington 1977; in press). A notable difference is, however, the absence of *Porcellispora longdonensis* in Cotham Beds assemblages from the Cliff Farm Borehole though this taxon is recorded from that facies in the above-mentioned Oxfordshire sections. Also noteworthy are the organic-walled microplankton associations, indicative of a marine environment or, at least, one subject to a significant marine influence, in both the Cotham and Westbury beds of the Cliff Farm Borehole. The presence of such remains is compatible with the view that the Westbury Beds are of marine origin but the belief that the Cotham Beds formed in a lagoonal or freshwater environment requires modification in the light of the similarity of the organic-walled microplankton assemblages of that facies to those from the more obviously marine Westbury Beds.

The palynomorph assemblages from the Cliff Farm Borehole may be regarded as late Triassic (Rhaetian) in age though, despite detailed palynological work on Alpine late Triassic deposits (Morbey 1975), an acceptable palynologically documented Rhaetian stratotype does not yet exist and the precise correlation of British Rhaetic deposits with the classical European late Triassic stages remains potentially liable to revision.

3. Samples from the Keuper Marl

A total of 76 Keuper Marl samples were examined from the following units and depths in fifteen boreholes sited along the Rhaetic and basal Liassic outcrop between Wapley Common and Bitton (listed below from north to south).

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Borehole ST 78 SW 13, Chescombe Farm, Wapley (p.63)
TGM: 7.90 m, 8.30-8.32m, 8.50-8.53m, 9.00m, 9.50-9.53m, 10.22-10.26m, 11.00 m.
Keuper Marl (undifferentiated): 12.50 m, 16.96-17.00 m.

Borehole ST 78 SW 14, Cliff Farm, Wapley (p.64)
Keuper Marl (mudstone with SEB): 35.72-35.75 m, 35.98-36.02 m, 36.48-36.53 m, 36.72-36.75 m.

Borehole ST 78 SW 12, Cliff Farm, Wapley (p. 62)
TGM: 7A7-7.81 m, 9.20-9.25 m, 9.46-9.50 m.
Keuper Marl (above SEB): 10.02 m, 13.10-13.13 m, 15.72 m, 22.46-22.50 m, 22.97 m.
Keuper Marl (SEB): 23.96-23.99 m.
Keuper Marl (below SEB): 24.48-24.52 m, 27.28 m.

Borehole ST 77 NW 22, Westerleigh Hill (p. 61)
Keuper Marl (above SEB): 15.44 m, 22.94-22.97 m, 25.48-25.53 m, 26.40 m.

Borehole ST 67 NE 56, Cliff Farm, Westerleigh (p. 49)
TGM: 10.65-10.70 m.
Keuper Marl (above SEB): 19.29-19.31 m, 22.34-22.36 m.

Borehole ST 67 NE 60, Leigh Farm, Pucklechurch (p. 53) TGM: 8.40 m, 9.20 m, 10.90 m.

Borehole ST 67 NE 59, Park Farm, Pucklechurch (p. 52)
TGM: 8.50 m, 9.00 m, 9.99 m, 10.50 m, 11.00 m, 11.50 m.
Keuper Marl (above SEB): 12.00 m, 18.25 m.

Borehole ST 67 NE 57, Coxgrove Hill, Pucklechurch (p. 50)
Keuper Marl (above SEB): 13.04-13.06 m, 16.44-16.45 m, 17.07-17. 17 m.

Borehole ST 67 NE 62, Gingells Farm, Shortwood Hill (p. 55)
Keuper Marl (above SEB): 16.00 m, 16.71 m, 18.00 m.
Keuper Marl (below SEB): 19.50 m.

Borehole ST 67 NE 61, near Gingells Farm, Shortwood Hill [p. 54)
TGM: 5.50 m, 6.50 m, 7.50 m.
Keuper Marl (mudstone with SEB): 13.25 m, 13.70 m, 14.00 m, 14.50 m, 15.00 m, 15.40m.
Keuper Marl (below SEB): approx 23.60 m.

Borehole ST 67 NE 58, Lodge Farm, Siston (p. 51)
Keuper Marl (mudstone with SEB): 14.60-15.29 m.

Borehole ST 67 NE 63, Saint Anne's Well, Siston (p.56)
Keuper Marl (above SEB): 7.94 m, 11.24-11.44 m.
Keuper Marl (below SEB): 23.99-24.24 m, 33.85 m.

Borehole ST 67 SE 24, Blue Lodge, Wick (p.58)
Keuper Marl (above SEB): 18.25-18.28 m.

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Borehole ST 67 SE 25, Redfield Hill, Oldland Common (p. 59)
Keuper Marl (above SEB); 12.75-12.80 m, 13.00-13.05 m, 13.20-13.25 m, 13.50-13.55 m, 13.75-13.80 m, 14.00-14.05 m.

Borehole ST 67 SE 26, Upper Cullyhall Farm, Oldland (p.60)
TGM: 8.93-9.13 m.
Keuper Marl(SEB): 17.31-17.34 m, 19.77-19.87 m, 20.02-20.16m,20.59-20.74 m.

Only one of the above samples (that from 22.3422.36 m in Borehole ST 67 NE 56, Cliff Farm, Westerleigh) yielded determinable miospores, However, the taxon recorded (*Classopollis sp*.) only serves to indicate a maximum age of late Triassic for the sample concerned.

Two samples of mudstone from Triassic infillings of karst fissures in Carboniferous Limestone at Slickstones Quarry (ST 700 917), near Cromhall, Gloucestershire, (Robinson 1957) were also examined in connection with the celestite study but were found to be devoid of palynomorphs.

The almost total absence of palynomorphs from the upper part of the Keuper Marl in the Cromhall - Yate - Bitton district is, in the writer's experience, fairly typical of that unit throughout Britain. Exceptions occur however, palynomorphs having been recorded from the Tea Green Marl in Nottinghamshire (Morbey 1975), Dorset (Orbell 1973) and from the Grey Marl facies of the highest Keuper Marl of Glamorgan (Orbell 1973), Somerset (Warrington 1974) and south Devon (Stevenson & Warrington 1973).

ACKNOWLEDGEMENT. The author is grateful to Mr. E.F.P. Nickless for his helpful comments upon an initial draft of this contribution.

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THE CORNUBIAN ISLAND

by
E.M. Durrance and P. Grainger

Abstract The distribution of Permian, Jurassic and Cretaceous rocks in the area of South-west England suggests the presence of a Cornubian Upland in Permo-Triassic times which continued as a Cornubian Island during the remaining Mesozoic. But the absence of shore line facies particularly within the Lower Liassic, except in a few localised areas, has cast doubt upon this idea. The geotechnical properties of the lutites of the Littleham Mudstone Formation of the Permo-Triassic at Exmouth show that an overburden thickness of between 1000 and 1300 m of strata was at one time present. Of this thickness only 150-450 m represents rocks of Jurassic or younger age. It is suggested that about half of this was of Lower Liassic age. The development of the fault controlled differential subsidence which acted during Permo-Triassic sedimentation is thus considered to have passed into a period of general subsidence in the Lower Liassic. This may have caused the Cornubian Island to be covered. Differential movements must have continued, however, but on a local scale. Major differential movement probably started again at the end of the Lower Liassic, after which the Cornubian Island remained in existence until it was covered in the Senonian.

The occurrence of Permo-Triassic rocks in South-West England is largely restricted to the area east of a sinuous line running roughly north-south through Exeter, although a major westward extension occurs in the Crediton trough and a small area of Permian rock is found on the southern margin of Barnstaple Bay (Edmonds et al, 1975). Off-shore, however, Permo-Triassic rocks are much more extensive, largely surrounding the rocks of Devonian and Carboniferous age which form the land area (Avedick, 1975). A marked unconformity separates the gentle structures of the continental facies rocks of the Permo-Triassic from the highly deformed Devonian and Carboniferous beds. Although showing some overlap, because the Jurassic rocks of South-West England follow conformably on the Triassic their outcrop too is restricted. In this case to the area east of a sinuous line passing roughly north-south through Taunton. A marked westerly extension of the Jurassic outcrop occurs, however, along the Somerset coast. This extension is related to the important development of Jurassic beds beneath the Bristol Channel (Lloyd et al, 1973; Brooks and James, 1975), and

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reflects their general development in the Western Approaches (Avedick, 1975). The distribution of the Permo-Triassic and Jurassic strata in this fashion has led to the idea that South-West England was an upland area principally suffering erosion in Permo-Triassic times, which continued in positive relief as an island throughout the Jurassic. Indeed from the sequences of Cretaceous strata preserved both on and off-shore in the area it appears that this Cornubian Island continued its existence, but with varying size, until it became submerged during the Senonian stage of the Upper Cretaceous (Durrance and Hamblin, 1969).

Objections to this concept of a Cornubian Island have recently been raised by Owen (1976), who notes that although coarse marginal facies of Liassic age are found adjacent to Jurassic islands in the Vale of Glamorgan, Avon and Somerset, there is no indication of nearby land in the Liassic facies of Mochras, in the Jurassic of the Bristol Channel (except for a sand influx in the Oxfordian) or even in the Lower Liassic of west Dorset. The possibility therefore arises that much of South-West England (and Wales) may have been a subsiding area throughout the Jurassic, with the influx of, for example, the Bridport Sands, possibly resulting only from periodic uplift and the erosion of parts of earlier deposited sequences. The very thick developments of Jurassic rocks which occur in the basins mantling the Cornubian Island (Whittaker, 1975) then become zones of even more rapid subsidence, probably faulted in graben. In broad terms this picture of differential subsidence continued during the Cretaceous, but an important period of deformation and erosion occurred before the deposition of the Cretaceous Upper Greensand and Chalk.

The determination of the maximum thickness of overburden experienced by shaley rocks, and the effects that increasing overburden pressures have on physical properties of shale have recently been investigated by Magara (1976). From a study of the Cretaceous shale of western Canada he has obtained the following empirical relationship: $Ø = 0.466\ {}_{\Delta}t - 31.7$

where Ø is the in situ porosity in percent and ${}_{\Delta}t$ is the sonic transit time in microseconds per foot. Studies of the red lutites in the Littleham Mudstone Formation of the Permo-Triassic series near Exmouth, Devon, have shown that these rocks have an in situ porosity of about 25% and possess a some transit time of about 120 microseconds per foot. Thus although these are a non-marine deposit without shaley

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partings, the relationship appears to hold true. Using the method of Magara (1976) to relate porosity to overburden pressure, a thickness of about 1300 m of rock of normal density is obtained.

Estimates of the amount of overburden can also be obtained from consolidation tests (Altscheffl and Harrison, 1959). Using an oedometer with a maximum applied pressure of 3200 kNm^{-2}, plots of void ratio against effective pressure for two undisturbed and three remoulded samples of the same lutites from the Littleham Mudstone Formation show a convergence between the extrapolated curves of the undisturbed and remoulded samples at a pressure of about 2500 kNm^{-2}. This is equivalent to an overburden thickness of about 1000 m of rock of normal density, Because of the remoulding technique employed, and uncertainties about when the reddening of the material occurred during its consolidation history, this value must be regarded as a minimum value. Combining the results of these consolidation tests with those from the Magara formula produces a result for the overburden thickness with minimum and maximum limits of 1000 and 1300 m respectively.

The thickness of the Permo-Triassic Series above the sampled · levels of the Littleham Mudstone Formation is shown by Henson (1971) to be of the order of 850 m, This indicates that the maximum post-Triassic cover experienced in the Exmouth area was of the order of 450m, but this cover may have been as little as 150 m. The average value of 300 m would not seem unreasonable. In particular, this thickness is restricted to any Jurassic to lowermost Cretaceous (Wealden) rocks which may once have been present in the area, as deformation and erosion caused the removal of even the greater part of the Triassic prior to the deposition of the relatively thin Cretaceous Upper Greensand and Chalk. Moreover, while comparing closely with thicknesses of about the same value for the Liassic of west Dorset given by Arkell (1933) and Rayner (1967), this is less than the thickness of about 350 m given by Whittaker (1975) for just the Lower Liassic in the Somerset basin.

It is not possible to indicate precisely which parts of the Jurassic to lowermost Cretaceous succession were once present in the Exmouth area, but the general characteristics of the facies of rocks within this age range which lie to the east, suggest that the Lower Liassic is likely to have contributed about half of the sequence. Perhaps, therefore, the post-Hercynian rifts which so clearly developed in southern Britain during Permo-Triassic times became

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relatively quiescent in the Lower Liassic. During this time a more general subsidence was dominant, possibly resulting in the covering of the Cornubian island, with differential movements taking place only locally. This pattern probably did not extend beyond Lower Liassic times, since when more pronounced differential movements occurred and only intermittently were beds deposited on the Cornubian Island itself.

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TURONIAN MICROBIOSTRATIGRAPHY OF BEER, S.E. DEVON

by Malcolm B. Hart & Phillip P.E. Weaver

Abstract The Turonian succession at Beer, S.E. Devon, has been investigated for its microfaunal content, and the distribution of the Foraminifera and Ostracoda described in terms of general palaeo-oceanographic changes. The rich, though low diversity, planktonic foraminiferal fauna can be used to recognise three main assemblages, viz. *Globotruncana imbricata* MORNOD / *G. marginata* (REUSS), *G. sigali* REICHEL/ *Praeglobotruncana helvetica* BOLLI, and *G. pseudolinneiana* PESSAGNO/ *G. coronata* BOLLI. The benthonic microfauna agrees with this outline sub-division, the central unit of which appears to be a relatively deeper water fauna.

1. Introduction

The Turonian succession at Beer, S.E. Devon, was described by Meyer (1874), Jukes-Browne & Hill (1904), Rowe (1903), and Owen (1970). More recently it was mentioned by Kennedy & Garrison (1975) and Hancock (1975) in general descriptions of chalk petrology. The Turonian at Beer is a particularly interesting succession from a sedimentological standpoint, as it shows several hardgrounds and nodular horizons, probably suggesting deposition in a shallower part of the chalk sea (Hancock 1975).

The Turonian, as defined by Rowe (1903), included the zones of *Rhynchonella cuvieri*, *Terebratutina gracilis*, and *Holaster planus*. However, there has been much debate in recent years as to where the lower and upper boundaries of the stage should be placed (e.g. Kennedy & Garrison 1975; Hancock 1975; and Hart 1975). For the purpose of this paper the Turonian encompasses the zones of *Inoceramus labiatus* (cf. *R. cuvier*), *T. lata* (cf. *T. gracilis*) and *H. planus*. Disputed horizons are, in the main, not well-developed at Beer, or are unprocessable as indicated in Fig. 1. Horizon 'A' of Kennedy & Juignet (1973) is not recognisable micropalaeontogically in the Beer succession; therefore, for convenience, all the chalk above the Cenomanian limestones is regarded as Turonian.

Samples taken at approximately two metre intervals were examined for Foraminifera and Ostracoda. In general, good, well-preserved faunas were obtained, except in the lower levels where the

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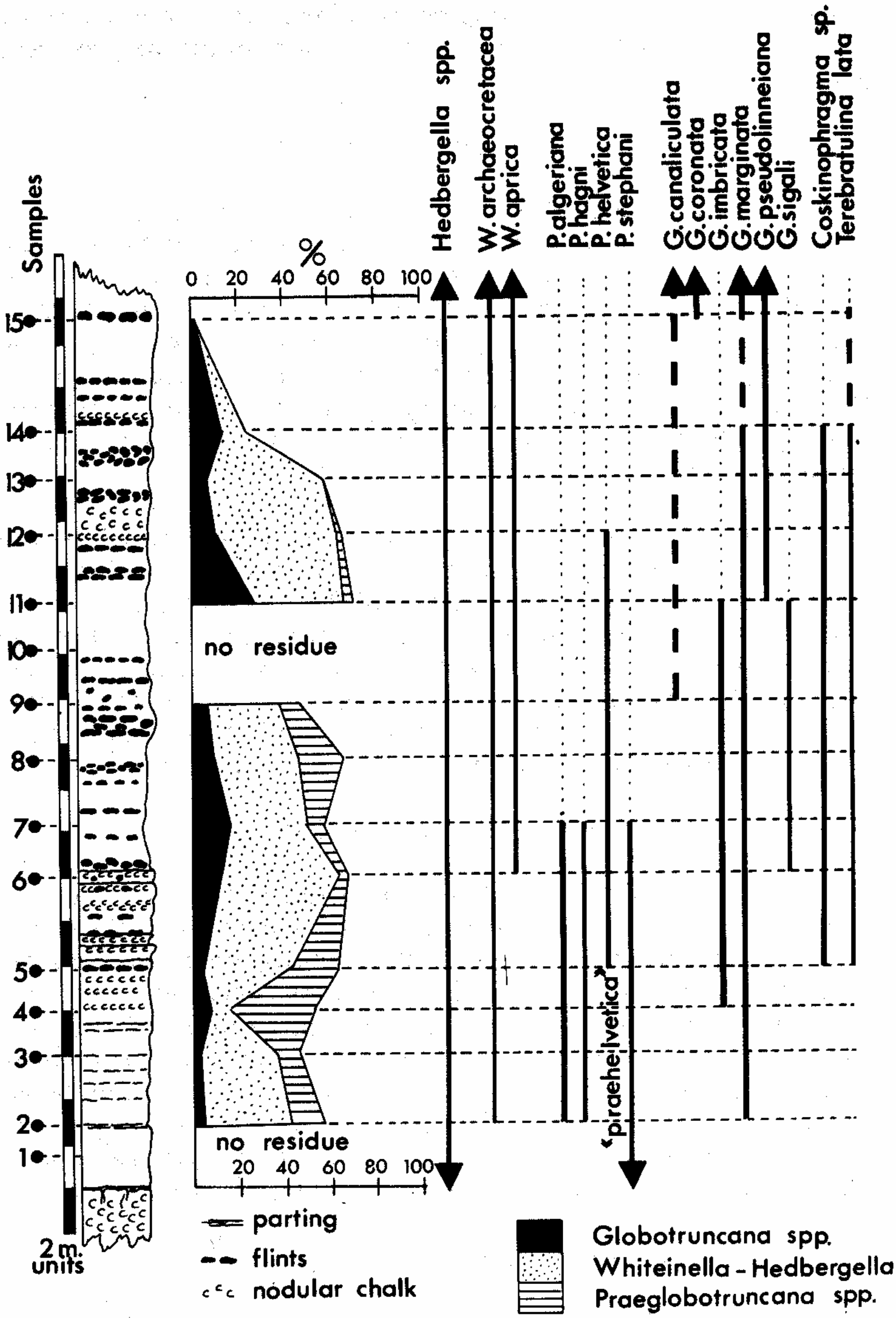

FIGURE 1
Lithological and palaeontological data for the Beer Harbour cliff section (east side), S.E. Devon.

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chalk was too hard. Broken fragments of inoceramids and *T. lata* are fairly common and make up substantial parts of all the processed residues.

2. Lithological succession

In total, some 37 metres of Turonian were measured and sampled (Fig. 1). The base of the section is marked by the non-sequential junction with the Cenomanian limestones. The immediately overlying 'Turonian' consists of massive, tough, gritty (due largely to broken inoceramids) chalks which are nodular and contain hardgrounds, suggestive of a condensed sequence. Phosphates are common, and many nodules have phosphatic coatings. This basal unit is approximately 10 m. thick, and extends up to the first level of flints.

One metre above the first flint horizon are two distinctive hardgrounds and about two metres above is another pair. These are the 'two pairs of yellow nodular bands' described by Rowe (1903). Above this level true hardgrounds are not found, although there are several nodular horizons. In the remainder of the succession the chalk contains many distinct flint bands as well as layers of scattered flints. The 'two foot band' and 'four foot band' of Rowe are clearly visible in the section as beds of marly chalk without flints.

3. Microfaunal analysis

The foraminiferal fauna, characteristically for southern England, is limited to only a few species. Unlike the rest of the British mid-Upper Cretaceous succession, the planktonic Foraminifera are very abundant. Fig. I shows the planktonic/benthonic ratio (for discussion of the use of this ratio see Hart and Carter 1975) for the 250-500 size fraction. From this it can be seen that there is an average planktonic fauna of some 50-60% of the total assemblage, which becomes less than 20% only in the uppermost sample. The benthonic fauna is limited to relatively few, long-ranging species, the majority of which have only limited stratigraphic value.

1. The planktonic Foraminifera

The planktonic Foraminifera are dominated by the *Whiteinella/Hedbergella* plexus. In general terms *Whiteinella* seems, as a generic group, to evolve from the *Hedbergella* lineage in

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the vicinity of the Cenomanian/Turonian boundary, although the relationships are by no means clear at the moment. This transition is better seen in more southerly areas (e.g. S.E. France), and the poor record seen in the U.K., coupled with the poor preservation in the lower levels of the Beer succession, makes any interpretation of the fauna highly tentative. The greater part of the Turonian is, however, dominated by *W. archaeocretacea* PESSAGNO, which is joined, higher in the succession, by *W. aprica* (LOEBLICH & TAPPAN).

The *Praeglobotruncana* fauna is represented in the lower part of the succession by the *P. hagni* SCHEIBNEROVA - *P. algeriana* CARON lineage, with the typically Upper Cenomanian species *P. stephani* (GANDOLFI) comprising only a minor part of the fauna. Above sample 7 (in the basal T. lata zone) the genus is represented only by *P. helvetica* (BOLLI), which is an important, world-wide, Turonian marker. The first occurrence of *P. helvetica* should mark an important biohorizon, although its appearance in this section is probably unreliable because the chalk at this level is so difficult to process. Beer also lies well towards the northern limit of the species' distribution, and as such, the first appearance is probably a migration event and not a true evolutionary event. The matter is further complicated by the occurrence in this section of *Hedbergella (?)praehelvetica* TRUJILLO, the ancestral form. There is currently little agreement as to the precise separation of these two species.

The *Globotruncana* fauna falls into three main units. The lower, characterised by *G. marginata* (REUSS) and *G.* cf. *imbricata* MORNOD, is poorly represented. This is partly due to processing problems, but such rare specimens as do occur are also totally overshadowed by the dominant *Praeglobotruncana* and *Whiteinella* faunas. Just above the first flint horizon and hardgrounds associated with the base of the *T. lata* zone, *G. sigali* and *P, helvetica* are then both important elements of the fauna until the levels of samples 11 and 12 respectively, where the two species are last seen. At about this level the third *Globotruncana* fauna appears, with species like *G. pseudolinneiana* PESSAGNO, *G. canaliculata* (REUSS) and *G. coronata* (BOLLI). The whole fauna of the topmost sample in the section is different from those below, and it is reminiscent of the fauna described by Hart & Carter (1975), Bailey (1975) and Carter & Hart (1977) from the uppermost Turonian and the Coniacian elsewhere in southern England.

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The overall distribution of the planktonic Foraminifera agrees quite well with that given in the account of the Cretaceous (Hart & Carter 1975) of southern England, with a high percentage of planktonic individuals throughout the greater part of the Turonian, reducing drastically close to the Turonian/Coniacian boundary. The *H. planus* zone (Bailey 1975; Bailey pers. comm.) is characterised by a poor planktonic fauna dominated by *G. pseudolinneiana*. This species was recorded from the Beer section by Owen (1970) from within the *T. lata* zone (the *H. planus* zone not being collected in the course of his work). Sample number 15 must therefore be close to the base of the *H. planus* zone or possibly even within that zone. The lithological base of the zone as used by Rowe (1903) is close to the level from which sample 15 was collected, when traced by eye through grass and vegetation at the top of the cliff. The occurrence of *T. lata* in sample 15 is of doubtful significance as it is known to extend beyond the limits of the formally designated zone of its name.

2. the benthonic Foraminifera

The benthonic fauna is poorly represented throughout the succession but also falls into three distinct units, which nearly coincide with those already described for the *Globotruncana* fauna.

The lower part of the succession (cf. *I. labiatus* zone) is characterised by the occurrence of *Arenobuliminapreslii* (REUSS), *Tritaxia tricarinata* REUSS, *Gavelinella berthelini* (KELLER) and *Lingulogavelinella globosa* (BROTZEN), while the uppermost association of *A. preslii*, *T. tricarinata*, and *Gavelinella emscheriana* HOFKER is restricted to the topmost sample. The intervening association, which seems to coincide almost with the *T. lata* zone, and overlaps to a large extent with the *G. sigali/P. helvetica* assemblage, contains a poor, reduced, benthonic fauna. *Lenticulina spp., G. emschriana, Marsonella oxycona* (REUSS), and T. tricarinata are the main faunal constituents. Of particular interest, however, is the occurrence of large numbers (especially in the larger size fractions) of *Coskinophragma* sp. Bandy suggested (1960) that large, agglutinated, labyrinthic foraminiferids (such as *Coskinophragma*) are characteristic of outer shelf and/or bathyal depths (c. L150-200 m.). The distinct parallelism of the *Coskinophragma* fauna with *T. lata*, and the highest percentages of planktonic individuals would seem to confirm this suggestion. This postulated deeper water interval during the Turonian at Beer also appears to agree with the field evidence of nodular chalk with

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hardgrounds in both the upper and lower levels of the succession, separated by much Purer chalk with courses of flints. Bailey (1975), in his work on the Annis' Knob succession recorded neither *T. lata* nor *Coskinophragma* sp. in any of the nodular chalks with hardgrounds that characterize the upper levels of the *H. planus* zone and the lower levels of the *M. cortestudinarium* zone. The planktonic fauna from these higher levels in the Beer succession (0 - 17.5% from Bailey 1975, fig. 1) agrees with the above model.

Such a shallowing in the late Turonian/Coniacian has been suggested in several recent works (Kennedy & Juignet 1974; Juignet & Kennedy 1974; Hart & Carter 1975; Hancock 1975; Hart 1976) and must clearly be responses to the same event.

3. the Ostracoda

The occurrence of Ostracoda is largely related to the way in which the samples were processed. The hard samples from the lower levels of the section are poor in Ostracoda, whilst the upper samples all have abundant ostracod faunas. Turonian Ostracoda from Britain have not been studied in detail since the work of Jones & Hinde (1890), and hence many of the species obtained from Beer have not yet been described and named. The Ostracoda fall into twenty-one genera *Amphicytherura, Bairdia, Curfsina (.9), Cythereis, Cytherella, Cytherelloidea, Cytheropteron (Aversovalva), Imhotepia, Macrocypris, Monocertina, Karsteneis, Neochythere, Oertlieila, Orthonotacythere, Paracypris, Pedicythere, Pontocypreila, Pterygocythereis, Spinoleberis, Trachyleberidea, and Xestoleberis. Cytherella ovata* ROEMER, *Cytherelloidea hindei* KAYE, and *Imhotepia gr. marssoni* (BONNEMA) are species which occur throughout the section, whilst *Trachyleberidea acutiloba* (MARSSON) occurs in the upper six metres only and *Cytherella granulosa* (JONES) occurs only in sample 13. These last two species are normally recorded from higher in the succession, but it appears that their ranges can be extended back at least as far as the upper levels of the *T. lata* zone.

4. Conclusions

The Beer fauna is dominated by abundant planktonic Foraminfera which have a distribution similar to those species found elsewhere in southern England. At present the varied Ostracoda fauna cannot be compared in detail with other localities because of the lack of taxonomic data. It is possible, however, to recognise a

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deeper-water phase in the mid- to late Turonian because of the association of a rich planktonic fauna, a low diversity benthonic fauna, and the occurrence of *Coskinophragma*. This same species (as *C. irregularis* (ROEMER)) is also recorded from the mid-Upper Turonian of the Cap Blanc Nez section in Northern France (Robaszynski pers. comm.).

ACKNOWLEDGEMENTS. The authors wish to thank D.J. Carter, H.W. Bailey, C.J. Wood and F. Robaszynski for discussing this work on various occasions, and acknowledge the receipt of financial assistance from the Natural Environment Research Council (PPEW) and Plymouth Polytechnic Staff Research Fund (MBH).

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GLAUCONITE IN CELTIC SEA SEDIMENTS

by M. George and J.W. Murray

Abstract. Mineralogical glauconite occurs in the surface sediments of the Celtic Sea as loose grains and as infilling in the shells of the foraminiferids, gastropods, bryozoans, and in borings in the shell debris. The colour ranges from pale to dark green and appears to be correlated with the degree of ordering. Only the darker, more ordered varieties are found beneath the sediment surface. It is concluded that glauconite is forming in the area at present and that the darkening and associated ordering takes place over a relatively short time period.

1. Introduction

The term glauconite has been used in a loose way to describe green sedimentary grains (morphological glauconite or glauconie of French authors) and the hydrated iron-rich clay mineral (mineralogical glauconite) (McRae 1972). Not all morphological glauconite is composed of mineral glauconite. Identification of the mineral is carried out by X-ray diffraction.

In the course of studying the recent Foraminifera, the occurrence of green grains in the sediments of the Celtic Sea was noted. The present contribution records the investigation of the mode of occurrence, geographical distribution and mineralogy of this material.

2. Description of the morphological glauconite

Sixty two sediment samples were examined from the area between Long. 7^{o} and 8^{o}30' W and Lat. 50^{o} and 51^{o} 35' N. The samples were collected from the top 1 cm. of the sediment, using a specially constructed grab. Each was washed on a 63 μ m aperture sieve, dried at 800C, and examined under a stereoscopic binocular microscope.

Morphological glauconite occurs in sixty of the surface sediment samples examined. Several different morphological

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forms are present and the colour ranges from pale green to emerald to dark green or almost black. The following types occur.

Grains. Much of the morphological glauconite is present as irregular shaped grains of varying size. The dark green grains are of the form described by Triplehorn (1966) as mammillated and lobate pellets, i.e., they are irregular, bear rounded knobs and are cut by cracks which are commonly infilled with white clay minerals. Some of the emerald grains have this form too. The pale green grains are normally smooth and not cracked and they have the shape of the chamber infilling of a foraminiferid or gastropod.

Infillings. The morphological glauconite infilling cavities in bioclastic material is commonly pale green although some emerald and dark green examples have been observed. There are three main modes of occurrence:

(a) infilling the chambers of foraminiferids, especially miliolids but also in *Gaudryina rudis* Wright, *Ammonia beccari* (Linné) *Elphidium crispum* (Linné), and *Globorotalia truncatulinoides* (d'Orbigny). The mineral is visible because expansion of the clay during the glauconitisation process causes the cracking and breakage of the foraminiferid test. The pale green varieties show no cracks like those seen in the dark green grains.

(b) infilling borings cut in mollusc shell debris. It appears that after the borings were made, morphological glauconite infilled the cavities and subsequent, abrasion of the shell has exposed the glauconite. Pale and dark green infillings have been seen but not all borings are infilled with glauconite.

(c) infilling the chambers of gastropods, bryozoans or the stereome of echinoderm debris. In most examples the morphological glauconite is pale green but the dark green variety is sometimes present.

3. X-ray studies of glauconite

Samples of glauconite were chosen to represent light and dark types over as large a spread as possible of the surface ranges where each occurred. In three cases it proved possible to gather a light and a dark sample from the same locality.

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The amount of glauconite available in each sample for study was, in most eases, limited to a few grains. Powder-camera techniques were therefore employed.

The grains of glauconite were crushed between two glass slides and the powder mounted in a 0.3 mm. diameter Lindemann glass capillary. The sample was then placed in a Debye-Scherrer camera and surrounded by the film in asymmetrical (Straumanis) orientation.

Nine-hour exposures were made, using iron K α radiation at 540 watts on a Phillips PW 1010 X-ray generator.

It had been decided to categorise the glauconite on the criteria used by Bentor and Kastner (1965).

Class I - Mineral glauconite

(a) Well-ordered: IM: symmetrical and sharp diffraction lines at 10.1, 4.53α3.3 Å. Reflections ($11\bar{2}$) and (112) always present.

(b) Disordered: I Md: asymmetrical basal diffractions broadened at the base. Reflections ($11\bar{2}$) and (112) absent.

Class II - Interlayed glauconite

Basal spacing '$d_{(001)}$ >10.15 Å.

Study of the films resulting from the glauconite X-rays showed that slight broadening of all lines had occurred, probably as a consequence of small crystal size. It was therefore decided to obtain a value of basal spacing (d 001) by measuring the line-positions of all reflections and calculating the C-axis parameter of the unit cell by means of a standard computer regression program. The presence or absence of the (112) and (112) lines, and the nature of the basal (001) reflection, were assessed by inspection at the time of measurement.

It was found that the glauconite fell into two types. One, the dark, mamillated grains, showed a basal spacing in the range 10.0 - 10.19 Å, with the lines (112) and (112) diffuse but discernible and (001) line fairly well defined and symmetrical. Slight variation in the sharpness of the lines on the films from sample to sample could be noted; this may be a function of sample preparation or of crystallite size. On the basis of Bentor and Kastner's classification, all the dark glauconites studied from the Celtic Sea sediments were judged to be slightly disordered but still to lie in the well-ordered Class la group.

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The other, light type of glauconite showed basal spacings of 10.14 Å, and above, with neither (112) nor (112) reflections present, and the basal (001) reflection broad, diffuse and apparently asymmetrical. These were assigned to the Class II (interlayered) group of Bentor and Kastner.

Comparison tests were made upon three samples of glauconite from sedimentary rocks, one from the Cretaceous, Hibernian glauconite greensand, Gobbins, Island Magee, N. Ireland and two from the Tertiary Bracklesham Beds, Fisher Bed VII, Whitecliff Bay, Isle of Wight. The dark Cretaceous and darker Bracklesham glauconites, proved to be highly ordered, with basal spacings of 10.1 - 10.15 A and sharp, defined (112), (112) and (001) lines. An emerald green glauconite from the Bracklesham sample proved to be less well-ordered, similar, in fact, to the Celtic Sea dark glauconite.

4. Distribution

The area sampled varies in depth from 75 to 130 m. The surface sediments vary from clean washed, well sorted, quartz medium sands on Labadie Bank (Hamilton et al. 1974) and the tail of Haig Fras to poorly sorted, muddy to very muddy, fine and medium sands with some biogenic debris, over much of the area. Locally there are patches of pebbles and coarse sand.

Bouysse et al. (1976) noted that in late glacial times sea level fell to around -110 to -120 m. in the Celtic Sea. During the Flandrian transgression sedimentation was initially a conglomerate resting on an eroded surface of older sediments. Above the conglomerate were deposited sands and muddy sands as the water deepened (Belderson and Stride, 1966).

Dark glauconite grains are present in nearly all the surface sediment samples (Fig. 1) and in all types of sediment. Pale green grains are found in the muddy sediments but not in the clean sands on the banks. Their distribution is far less widespread than that of the dark grains and they are only patchily present in the west and southwest of the area.

Infillings of the various kinds are present only in the muddy sediments and mainly in the north eastern half of the area. Those in

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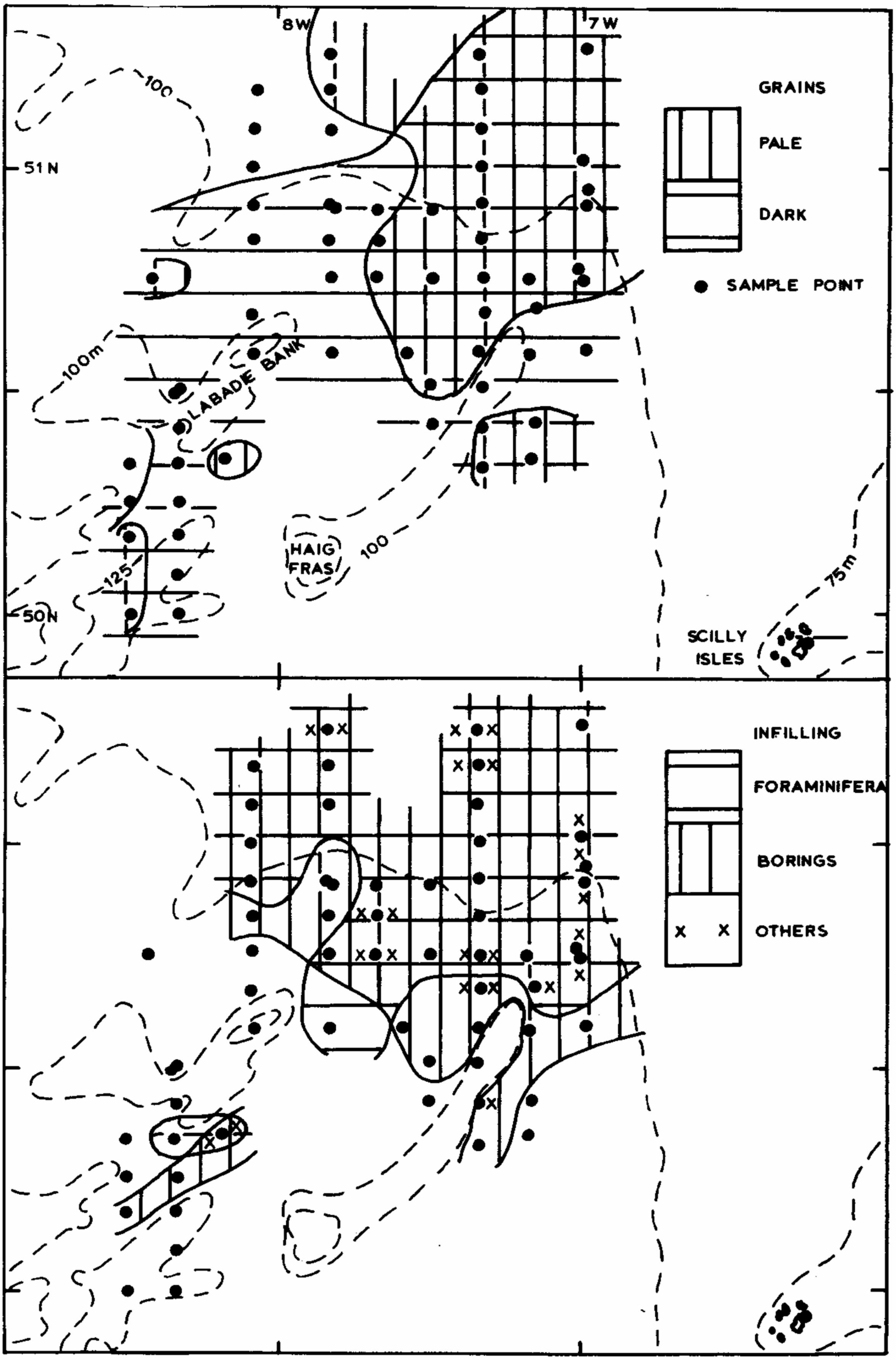

FIGURE 1. Distribution of glauconite in the surface sediments of the Celtic Sea.

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gastropods, echinoids, bryozoans, etc., are less widespread than those in foraminiferids and borings.

Two cores were studied to determine the distribution of glauconite within the sediment succession. One core, taken at Lat. 50^{o} 35' N, Long. 7^{o} 55' W, yielded a greenish muddy sand at its base, 67 cm. below the sea floor. This contained well-ordered dark glauconite grains and glauconite infilled borings in shell debris. The second core consisted of an upper graded sequence above a slightly muddy, greenish sand. The latter contained well-ordered emerald to dark grains and a foraminiferal assemblage of subrecent age. The graded unit (27 cms.) commences with a Turritella sand with pebbles overlain by muddy sand with shell debris. The latter yielded black grains, glauconite infilled borings in shells and pale glauconite in foraminiferids in a sample taken from the surface.

5. Discussion

Morphological glauconite has been recorded from the English Channel by Dangeard (1928), Murray (1965) and Clark (19703It has not previously been reported from the Celtic Sea and the results presented here are the first confirmation of the occurrence of mineralogical glauconite in British seas. It has, however, been reported from other parts of the European seaboard (Bay of Biscay, Caillère and Giresse 1966; off Spain, Lamboy 1968a and b, Odin and Lamboy 1975). A very immature glauconite has been recorded off Norway (Bjerkli and Ostrno-Saeter 1973).

The Celtic Sea glauconite shows no clear correlation with sediment type except for the absence of all except dark grains from the bank sands. There is no correlation with water depth. The thermocline extends down to around 50 m. and is only developed in the summer (Cooper 1967). The bottom water has a temperature range of 9.50 to 10.5°C. These values compare favourably with those reported from the areas of glauconite formation in the Niger Delta (Porrenga 1967) and off Gabon and Congo (Giresse and Odin 1973).

It is known from previous studies (see review by McRae, 1972) that mineralogical glauconite may form authigenically either by the alteration of biotite or from clay minerals. It may also be present in sediment as reworked detrital grains.

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The Celtic Sea glauconite is not thought to have been derived as an alteration product of biotite because of its widespread occurrence as infillings of shells. The Foraminifera, mollusc, echinoid, bryozoan and barnacle remains which contain glauconite infillings can all be matched with forms which are alive today. This gives conclusive proof that the pale green glauconite within biogenic material must be very recent in origin. Since the pale grains are clearly infillings which have separated from their host biogenic material these also must be of recent origin. It is believed that all this material has accumulated more or less *in situ*.

The majority of the dark grains are too large to have formed within borings, foraminiferids, or any organism except gastropods. Their irregular shape is not consistent with their having been infillings or faecal pellets. A few infillings have been found which are dark and in one instance a foraminiferid infilled with pale glauconite appeared to have been bored and the borings infilled with dark glauconite. This evidence suggests that dark glauconite is not necessarily very old even though it is more ordered than the pale form.

The evidence from glauconite distribution in the cores shows clearly that pale green, poorly ordered glauconite is present only in the surface sediments. At depth all the grains are emerald or dark and show good ordering.

All this evidence suggests that glauconite is forming in the Celtic Sea today as the pale green variety. This is poorly ordered but after a relatively short period of time it becomes more highly ordered and darkens to emerald or dark green.

ACKNOWLEDGEMENTS. We thank Mr. D. Hamilton of the University of Bristol for allowing us to see the two cores and, together with Dr. A.H.B. Stride of the Institute of Oceanographic Sciences for critical comment on the manuscript.

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POLYPHASE FOLDS FROM THE START COMPLEX

by D.M. Hobson

Abstract. The sequence of small-scale structures in the schists of the Start Complex consists of three generations of east-west trending folds, and is similar to the structural pattern in south Cornwall. The major structures comprise an F2 anticlinorium, together with large FI isoclines whose steeper limbs are replaced by tectonic slides. The Start boundary fault may be the eastern extension of the Perranporth -Mevagissey line.

1. Historical setting

The metamorphic rocks of the Start Complex are exposed around the southernmost tip of Devon (Fig. 1). They have always been something of an enigma, mainly because of the question raised during the last century of their ages: they could be either part of the local succession or a small slice of older basement. The Geological Survey map of the Complex was made by Ussher (1904), who concluded that it comprises Upper Palaeozoic rocks at a higher grade than those elsewhere in Devon. Recent structural work has supported Ussher's ideas (Phillips 1961; Marshall 1962, 1965), but definitive palaeontological evidence for the age of the Complex has never been found. The object of this note is to describe some of the local structures, and to discuss the position of the Start Complex within the Variscan fold belt.

In a classic paper, Tilley (1923) described the petrography of the Start rocks. He recognised two identical groups of grey pelitic schists, which are composed primarily of quartz and muscovite. Chlorite and albite arc occasionally major constituents, and there are also a considerable number of accessory minerals. Tilley also described a group of green schistose rocks, which he believed to be metamorphosed tuffs and basic lavas. He recognised two distinct green schist facies characterised by the assemblages chlorite-epidote-albite and hornblende-epidote-albite.

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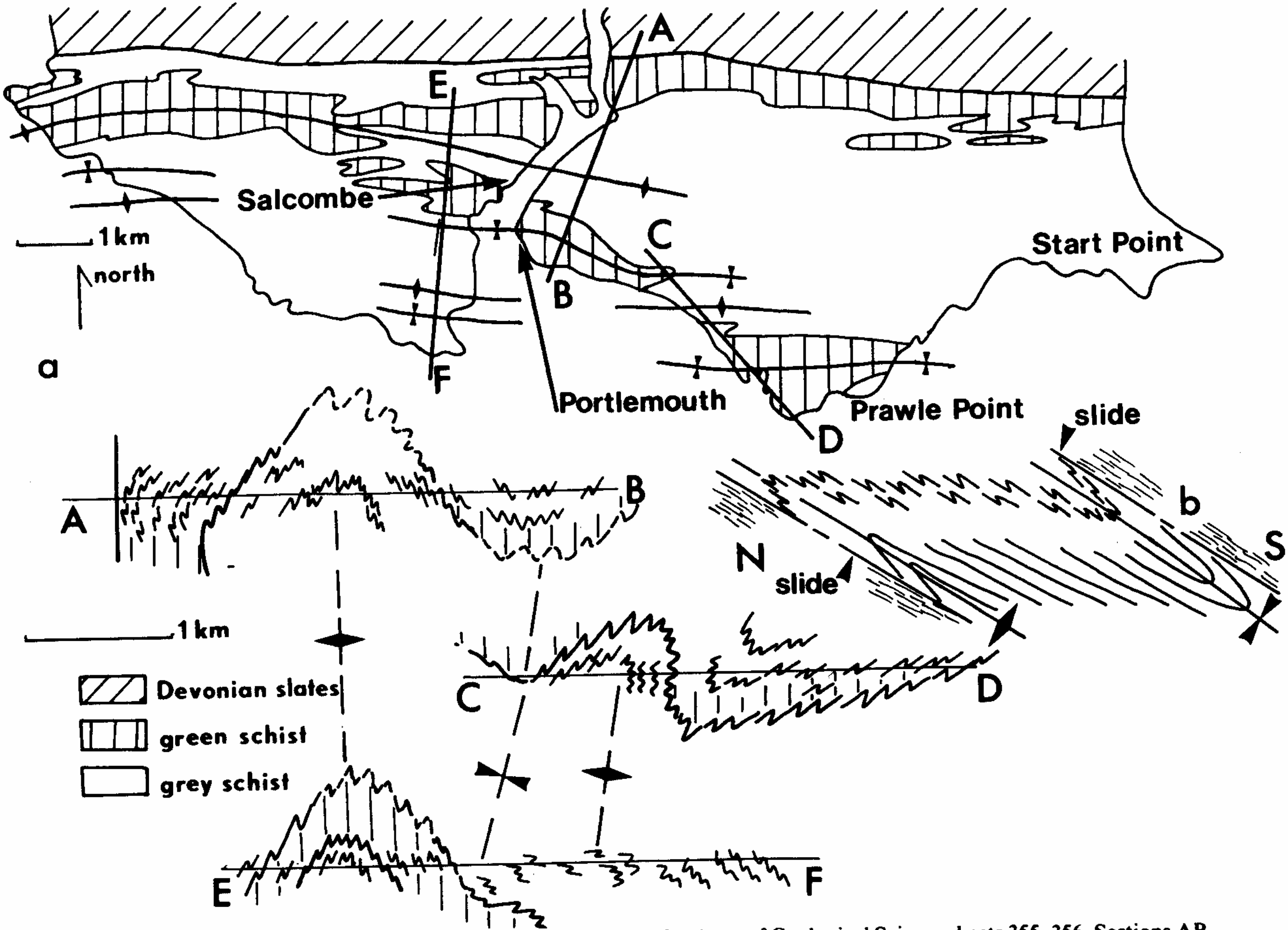

FIGURE 1. (a) Geological map of the Start Complex based on the Institute of Geological Science sheets 355, 356. Sections AB, CD and EF show the main F2 folds recognised by the author.
(b) Diagrammatic section through the green schists to show the shapes of small F1 folds and the position of inferred F1 isoclines.

2. Small-scale structures

Marshall (1962, 1965) recognised that the intricate geometrical patterns shown by the schists are the result of several phases of deformation. He described isoclinal Fl folds cut by subhorizontal schistosity. F2 folds are upright or slightly inclined and cut by a spaced crenulation cleavage. F3 and F4 folds are sets of kink bands, with steeply dipping and gently inclined axial surfaces respectively. Phillips (1961) noted that the schists are also characterised by a prominent lineation, which he believes is the result of microfolding of micaceous and chloritic layers.

Small Fl folds are only infrequently preserved in the Complex. Within the grey schists, folds of a later generation together with a penetrative lineation are prominent; only isolated hinge zones of FI isoclines are ever found. Within the green schists however, F1 folds are locally well developed, especially where there are few later formed structures. In particular, many F1 folds are visible in exposures of the green schists around the Salcombe estuary (Fig. 2). They are defined by layering, which is accentuated by the weathering of competent lithological units. Most of the Fl folds are not true isoclines in the sense of Fleuty (1964) because their interlimb angles lie between 50 and 250. The folds plunge towards the west, although there is considerable variation about the mean which may be a result of refolding. The attitude of their axial surfaces depends on their location on the flanks of later formed structures. Many of the folds exposed at Portlemouth (Fig. 2) east of Salcombe, are recumbent, because they lie on the gently inclined limb of a large F2 fold. Most of them are asymmetrical and the majority are overturned towards the north (Fig. 2b, c, d, 0. No direct way-up evidence has been found, and consequently the facing directions of the structures are not known.

Because layer boundaries are gradational, it is difficult to measure the precise shapes of these folds. Many of them are mixtures of Ramsay's (1967) class 1C and class 3 types (Fig.2a,b,d,). One limb is usually considerably attentuated compared with the hinge zone, and is a class 3 structure. The other limb is not so attentuated compared with the hinge zone, and is a class lC structure. One consequence of the shape formed by the interaction of class I C and class 3 folded layers is that the axial surfaces of the folds are nearly parallel to the more attentuated limb. The axial planes of these folds are therefore not the bisecting surfaces.

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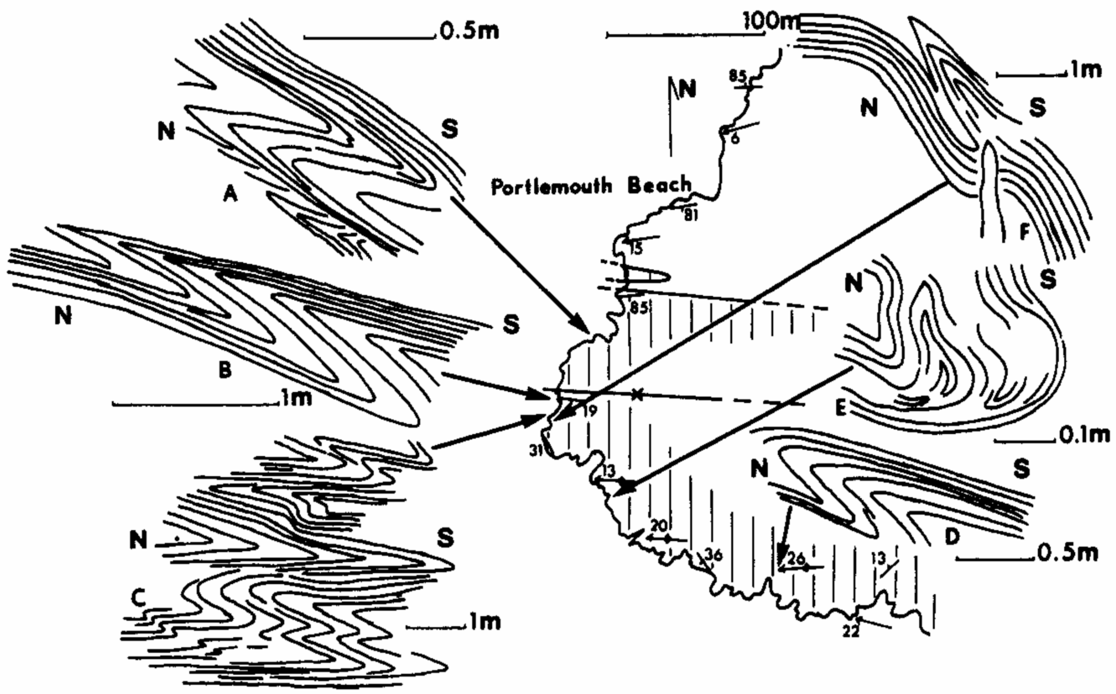

FIGURE 2. Small-scale folds from Portlemouth Beach, east of Salcombe.

Schistosity in these rocks (SI) is a mesoscopic planar fabric only in the grey schists, where it consists of aligned quartz and mica grains (Phillips 1961). Within the green schists, planar surfaces are not particularly obvious in hand specimen. There is however a locally prominent lineation formed of aligned prismatic hornblende and epidote grains, which is parallel to the axes of the Fl folds.

Small F2 folds are extremely common in the grey schists and are most easily seen where thin, discontinuous lenses of quartz, which formed parallel to the schistosity, are themselves folded. Phillips (1961) showed that these folds plunge towards the west and are parallel to a conspicuous crenulation lineation in the grey schists. Marshall (1962) believed that the axial planes of the F2 folds are almost vertical. The present author has found localities where the attitude of the F2 folds changes locally, from steeply inclined to almost recumbent. In general, small F2 folds with steeply dipping axial planes are present on the more gently inclined limbs of large F2 folds (Fig. I section CD). The small folds are frequently asymmetric and there is usually a consistent direction of overturning either towards the north or the south. On some more steeply inclined limbs of large F2 folds, small parasitic folds are extremely common, and their axial planes are gently inclined (Fig. I section CD). These folds are also asymmetrical, but there is often no recognisable consistent direction of overturning. Some of these small folds may have formed after the F2 deformation (see below).

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Within the grey schists, small F2 folds show a variety of shapes and vary from open to tight. Many of them are cut by a roughly axial planar crenulation cleavage (S2), in which individual planes are spaced at about 10mm. In thin section, euhedral grains of mica are formed along some cleavage surfaces. In the green schists, F2 folds are infrequent and are monoclinal in shape, with wavelengths in excess of I m (Fig. 2f).

Several sets of kink bands, recognised by Marshall (1965) as his F3 and F4 structures are sporadically formed throughout the Start Complex. There are also many recumbent folds of schistosity which are common on the flanks of upright F2 folds. These recumbent folds are not kink bands, because both limbs are of about the same length, and the hinge zones are rounded not angular. In a few localities these folds are cut by a flat lying crenulation cleavage (S3), which locally displaces the S2 fabric. This evidence suggests that the recumbent folds formed after the F2 event. Their relationship to Marshall's (1965) F3 and F4 folds is not clear

Where the axial planes of small F2 folds are gently inclined, it is impossible to distinguish them from the later generation recumbent folds. It is possible that the deformation which formed the recumbent folds also caused the rotation of some F2 folds into a nearly horizontal position. Interference between the two fold generations with almost identical geometries may account for the complex patterns visible at many grey schist localities.

Because small folds of about the same size but which belong to separate generations are locally developed together, refolded folds are present at some localities. Some examples of refolded Fl folds from the Salcombe area are shown in Figure 2 e & f. The interference patterns are of Ramsay's (1967) type 3, formed where the fold axes of successive generations are parallel. The refolded folds are most commonly found where the schistosity, S1, is steeply inclined and where the axial planes of the later phase folds are gently dipping. This may be due to the extreme frequency of small late-formed folds on the steep limbs of large F2 structures. Late generation folds are comparatively rare in the green schists, and consequently refolded folds are uncommon. One example has been seen at Portlemouth beach (Fig. 21), where a pair of F1 isoclines are folded by an F2 monocline.

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3. Large-scale structures

Tilley (1923) was the first person to describe the regional structure of the Start Complex. He suggested that there are two groups of grey schists separated by a single green schist formation, all folded into a westerly plunging anticlinorium. Marshall (1962) showed that the large anticlinorium is an F2 structure which deforms schistosity. Recent examination of the accessible coastal sections near Salcombe confirms the presence of a large F2 antiform, together with a series of major parasitic folds on its southern flank (Fig. 1).

Ussher (1904), Tilley (1923) and Marshall (1965) all found it impossible to distinguish the grey schists in the core of the antiform from those above the green schists. This led to speculation that the two groups of grey schist are the same formation, repeated by an earlier formed large isocline, with a core formed in the green schists. Changes in the direction of overturning of small F1 folds may be used to determine whether such an isocline is present. If the green schists are folded into a large isocline, small FI folds near the base and the top of the group should be overturned in opposite directions. Near the hinge zone of the isocline, small F1 folds will either show rapid changes in the direction of overturning, or else both limbs will be of equal length and the vergence indeterminate.

The sections through the green schists around Salcombe are ideal for a search for a large isocline. The cliffs at Portlemouth beach are cut through the lower parts of the green schist formation. Small F1 folds are found throughout this section. The whole of the green schist group is exposed in a section immediately south of Salcombe. Small FI folds are common in the upper parts of the green schists here, but they are not visible near the lower boundary.

On Portlemouth beach, the isoclines closest to the lower junction of the green schists lie about 20 m above the base. Of I 1 folds visible here, 5 are overturned towards the south, 2 to the north and 4 have almost equant limbs. In sections at the same structural level near Prawle Point, isolated isoclines overturned both to the north and also to the south occur. This evidence suggests that the lower sections of the green schists lie close to the hinge of a larger Fl fold (Fig. lb).

Within the main green schist formation, on both sides of the Salcombe estuary, there are more than 50 small Fl isoclines, all

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overturned towards the north. Here the layering dips towards the south, and the green schists therefore lie on the normal limb of a larger isoclinal antiform (Fig. I b). South of Salcombe, the northerly overturned folds can be traced to within 15 m of the upper junction of the green schists.

The upper junction of the green schists is most fully exposed south of Salcombe, although even here the actual margin is hidden beneath the sand. In the green schists closest to the upper junction, there are 7 small Fl folds, all overturned towards the south, evidence which indicates that the upper units of the green schists may lie close to the hinge of a larger FI fold.

Thus the available evidence suggests that the green schists lie predominantly on the normal limb of a southerly dipping isoclinal antiform. The hinge of this fold may have formed close to the present lower boundary of the green schists. The hinge of the complementary isoclinal synform may be formed close to the present upper margin of the green schists (Fig. lb). Because of the proximity of the cores of the folds to the grey schist - green schist boundaries, both these junctions must be interpreted as tectonic breaks. The lower junction is demonstrably parallel to cleavage, and involves the interbanding of grey and green schist lithologies. This evidence suggests that the junctions are tectonic slides, which probably formed at the same time as the FI folds, and which cut out the green schists that should lie on the inverted isocline limbs.

The predominant northwards direction of overturning of small Fl folds indicates that movement along the slides is also directed towards the north. The grey schists, now below the lower green schist boundary, are thus likely to have been originally structurally above the green schists. The grey schists now above the upper green schist boundary were probably originally structurally below the green schists. It is therefore likely that the two grey schist groups are separate formations. It is not possible to decide which is the older, because of the absence of primary way-up evidence.

4. The Start Complex within the Variscan fold belt

There has been considerable discussion as to the position of the Start Complex within the Variscan fold belt. The observation that the axes of folds in the Complex are parallel to those in proved

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Devonian rocks to the north has been taken as evidence that the deformation of the schists occurred during the Variscan orogeny. There is also radiometric evidence for an event in the schists at 305 my (Dodson & Rex 1971), during the Upper Carboniferous.

Correlation of separate fold phases across the Start boundary is difficult. Marshall (1962) equated the FI folds across the junction. Later however, he correlated the F2 folds in the Complex with the FI folds in the Devonian rocks, on the basis that both possess steeply inclined axial planes (Marshall 1965). Recent work suggests that this assumption is not correct. The Fl folds in the Devonian rocks are upright only because they are refolded by a late open antiform (Hobson 1976). Traced northwards, towards Plymouth, the F1 folds become less steeply inclined.

Thus in the absence of any contrary evidence it is logical to follow Marshall's (1962) description and to correlate F1 events across the Start boundary. The tightness and diminutive size of the Start Fl folds compared with their counterparts to the north probably reflects their formation at a deeper structural level. The association of the Start Fl folds with a schistose rather than a slaty cleavage fabric, reflects a higher grade of regional metamorphism, also probably indicative of a deeper structural level exposed in the Complex.

In the Complex, however, there are many late-formed recumbent folds of schistosity which have no counterparts in the Devonian succession to the north. Folds with such a style are known from south Cornwall (Smith 1965; Stone 1966; Dearman 1971). In fact, the sequence of small scale structures described by Smith (1965) from the Hayle area is identical to the pattern in the Complex. Near Hayle, isoclinal FI folds arc refolded by upright F2 folds; recumbent F3 folds are formed where the layering is steeply inclined. This may be evidence that the Start Complex is properly to be included with the structural units of south Cornwall, rather than to be related to the Devonian of south Devon.

Several points follow from this conclusion. Turner (1969) argued that the late recumbent folds in south Cornwall are restricted to areas above the granite batholith and are related to its emplacement. If these folds are indeed present in the Start Complex, beyond the southern margin of the batholith, such a conclusion is incorrect. It also follows that the Start boundary is the eastern

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extension of the important Perranporth - Mevagissey line, which records a major facies change in the Devonian rocks. Such a correlation is consistent with Dodson and Rex's (1971) recognition of dates of 290 - 315 my along a zone from Mevagissey to Start Point, and may record Upper Carboniferous movement along the boundary.

ACKNOWLEDGEMENTS. I would like to thank Drs. J.R. Graham and M.B. Hart for their critical comments on this paper.

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