Proceedings of the Ussher Society

Research into the geology and geomorphology of south-west England

Volume 6 Part 1 1984

Edited by G.M Power

The Ussher Society

Objects: To promote research into the geology and geomorphology of south-west England and the surrounding marine areas; to hold Annual Conferences at various places in South West England where those engaged in this research can meet formally to hear original contributions and progress reports and informally to effect personal contacts; to publish, proceedings of such Conferences or any other work which the Officers of the Society may deem suitable.

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Proceedings of the Ussher Society
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Papers

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Obituary

Samuel Crosbie Matthews
1936-1983

Crosbie Matthews, who died suddenly in Uppsala on the 5th May, 1983, was a founder member of the Ussher Society and a notable contributor to work in south-west England. From these early interests he achieved an enviable scientific reputation in a range of fields. He was born on the 23rd May, 1936 in the isolated moorland mining village of Muirkirk in Ayrshire where his father, who played an active and leading role in village affairs, was with the Ministry of Employment. After early years at the village school, Crosbie joined Crumnock Academy. His sister read Geology as a science subject with her arts degree at Glasgow University and started his geological interests with local collecting. Thus he read for Geology Honours when he, in turn, went to Glasgow University (1954-1958) mixing this interest with enthusiasms for cricket and golf. Then, in 1959, he moved to Bristol University for post graduate work for the Ph.D. (awarded in 1966) doing research on the St. Mellion area. He joined the staff of the department at Bristol in 1960 remaining there until he retired on medical grounds in 1982.

His first paper was published in our *Proceedings* in 1962. He led for us and others field trips which demonstrated the klippe at St. Mellion and also exhibited his then remarkable finds of conodonts and goniatites his papers on which, published in *Palaeontology* in 1969 and 1970 established a reputation in these fields of study. In the same journal in 1969 he published on conodonts from Chudleigh. His work on the Lower Carboniferous was later extended to Ireland and resulted in many publications alone or jointly with his friends at Trinity College and the Irish Geological Survey. Trinity College was a happy base for him for many years.

Again in *Palaeontology*, in 1973, he broke new ground with a description of lapworthellids from the Shropshire Cambrian and thus opened up his work and collaboration with Swedish and Russian friends on bizarre cone-shaped microfossils occurring in the late Pre-Cambrian and Cambrian. At the time of his death he was working at Uppsala with Stefan Bergstrom on these faunas and it was a happy, congenial and productive time for him and several papers are in course of publication.

He contributed also to regional tectonic and sedimentary problems of the Hercynian particularly on the alleged Variscan front (which he rejected) and on links between Caledonian and Variscan tectonics. His overview of facies in south-west England in the *Neues Jarhbuch* (1977) is particularly well-regarded as is his more recent attempt, with colleagues, to explain local Hercynian basins in terms of delamination phenomena (*Sedimentary Geology* 1983) and pull-apart basins.

His excellent skills as a linguist stood him in great stead in Scandinavia, Russia, Germany and Belgium where he often travelled. He lectured well in both French and German. He was Visiting Professor at Göttingen (1979 - 1980). The penetration and quality of his contributions were highly regarded on the Continent and did much to enhance the reputation of Bristol University. So geology has lost a scientist who had established a reputation for excellence and originality. His many friends, here and abroad, have lost a companion of great humour, wit and astuteness whose breadth of interest made his company such a delight leaving good memories to mellow sadness.

M. R. House

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The Scott Simpson Lecture

Read at the Annual Conference of the Ussher Society, January 1984

Variscan-Caledonian comparisons: late orogenic granites

JANET WATSON, F.R.S.
M. B. FOWLER
J. A. PLANT
P. R. SIMPSON

Watson, J., Fowler, M. B., Plant, J: A. and Simpson, P. R., 1984 Variscan-Caledonian comparisons: late orogenic granites. *Proceedings of the Ussher Society*, 6, 2-12.

The Variscan granites of south-west England form the high points of a batholith at least 200 km in length which seems to maintain a rather uniform composition down to depths of 10 km. The geochemical signature of these granites closely resembles that of the Caledonian Cairngorm batholith in the Grampian Highlands of Scotland. The crustal settings of these two batholiths are shown to have little in common, but the tectonic regimes in operation at the time of emplacement were broadly similar; in both provinces, compressive stresses ended and phase of limited extension associated with block movements coincided with the onset of granite magmatism. The geochemical similarities of the batholiths are thought to reflect similarities in the processes of magma genesis and emplacement. The ultimate source is thought to be metasomatised mantle rather than crustal material.

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1 Introduction: the significance of granite provinces

The granites of a single orogenic province commonly show a family resemblance to one another which extends beyond the details of their mineral and chemical compositions to the characters of their associated minor intrusions and mineral deposits. The Variscan granites of western and central Europe form such a family group, contrasting, on the one hand, with the granites of the European Caledonides and, on the other, with the sparse Alpine granites of the Mediterranean region. The sum of the distinctions between granite assemblages of different provinces may have been determined by a number of factors, notably the plate tectonic setting, the regional stress regime, the character of the underlying crust and that of the subjacent mantle. In this paper, we explore some of the controlling factors by reference to two groups of geochemically similar granites emplaced within two geologically dissimilar tectonic provinces: the Variscan granites of south-west England and the Cairngorm group of Caledonian granites in Scotland. Some of the work on which the paper is based was carried out as part of an EEC-sponsored study by the Institute of Geological Sciences (now the British Geological Survey) of the relationship between uranium provinces granite magmatism and uranium mineralization.

2 The geological settings

The granites of south-west England are in many ways typical members of the Variscan family and show the characteristic tin-tungsten-uranium mineralisation widely developed in western and central Europe. Geophysical evidence shows that the exposed granites are merely the high points on a much larger buried batholith with a length of at least 200 km, a breadth of about 40 km and a height of at least 10 km (Fig. 1). This Cornubian batholith was emplaced in an assemblage of Upper

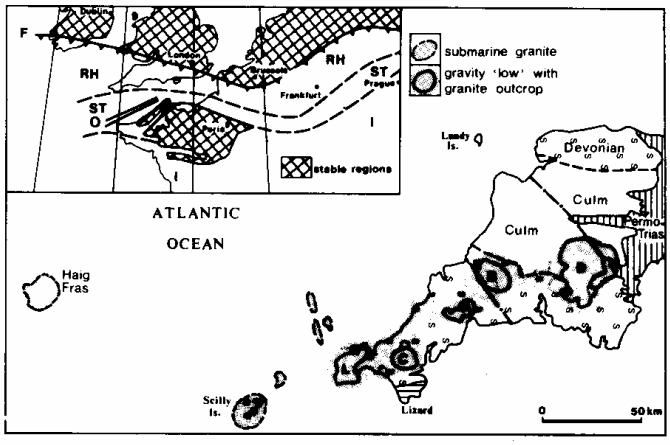

Figure 1. The distribution of Variscan granites in south-west England. The outline of the negative gravity anomaly corresponds to the zero isogal and shows the approximate extent of the buried Cornubian batholith. L = Land's End, C = Carnmenellis, A = St. Austell, B = Bodmin, D = Dartmoor granite.
Inset: the tectonic zones of the Variscan orogen. F = Variscan front, RH = Rhenohercynian, ST = Saxo-Thuringian I = Internal crystalline regions, O = Alderney-Ouessant line.

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Palaeozoic sedimentary and volcanic rocks which are strongly deformed but show, at most, only a low grade of metamorphism.

In terms of structural setting, the Cornubian batholith (with the nearby Haig Fras granite) is anomalous in that it is located in the Rhenohercynian tectonic zone, immediately adjacent to the northern foreland of the Variscan orogen, which is otherwise almost devoid of granites over a strike-length of more than 2,000 kin. Most of the granites of the Iberian peninsula, France and the Bohemian massif to which the Cornubian batholith may most readily be compared lie within the more southerly Saxothuringian tectonic zone or in the internal crystalline zone of the Variscan-Hercynian orogen (Fig. 1).

The Caledonian granites selected for comparison are those forming the east-west line of intrusions extending from the Monadhliath massif, through the Cairngorm and Lochnagar massifs to Mount Battock in the Grampian Highlands of Scotland (Fig. 3). A large negative Bouguer gravity anomaly centred on the granites shows that they are the high points of a buried batholith more than 100 km in length and some 10 km in height. This Cairngorm batholith is emplaced in the metamorphic zone of the Caledonides in which a strongly deformed and metamorphosed late Precambrian to early Palaeozoic (Dalradian) sedimentary and volcanic assemblage is tectonically interleaved with pre-Caledonian basement rocks of high metamorphic grade. The metamorphic province is penetrated by many late Caledonian granite plutons of much the same age as the Cairngorm batholith and the latter may therefore be said to be in context with respect to the structural setting. In terms of chemical composition, however, it appears strongly anomalous in that it is rich in a suite of trace elements normally represented at average or below average levels in the province.

The interest of the Cairngorm batholith in the context of this paper lies in the fact that the geochemical features that render it anomalous with respect to other Caledonian granites are matched with remarkable precision by those of the Cornubian batholith. We have therefore an opportunity to compare two batholiths which are located in orogenic provinces of differing ages and tectonic styles and which are each in some way anomalous with respect to their host provinces.

In the next sections of this paper we summarise the common features of the batholiths and examine the characters of the enclosing crustal complexes as a preliminary to a discussion of the factors influencing their formation.

3 Characteristic features of the batholiths

Preliminaries

In comparing the Cornubian batholith with that of Cairngorm, it is necessary to distinguish at the outset between the features determined during the primary stages of magma genesis, emplacement and crystallisation of magma on the one hand and those resulting from pneumatolytic and hydrothermal reactions on the other. The resemblances between the two batholiths (Table 1) are those connected with the primary stages. Conspicuous differences are apparent with respect to the late-magmatic and post-magmatic stages, the Cornubian batholith being associated with a varied assemblage of mineral deposits and the Cairngorm batholith being almost barren. It has been suggested elsewhere (Plant *et al.,* 1980, 1983) that these differences were connected with the extent and style of interaction between the granite and its envelope, a subject which is touched on in a later section. Here we will concentrate on the primary characteristics.

Analytical techniques

Major elements were determined on a Philips PW1212 x-ray fluorescence spectrometer at Imperial College, using the method of Parker (1982). Ba, Rb, Nb, Sr, Zr, were determined on a Philips PW1400 XRF at Bedford College and REEs, Th, Hr, Ta by INAA. U was analysed by the delayed neutron method.

TABLE 1: CHARACTERS OF THE BATHOLITHS

Cornubian Batholith	Cairngorm Batholith
> 200 km length	> 100 km length
c10 km thickness	c10 km thickness
low density	low density
high heat production	high heat production
pegmatites locally numerous	pegmatites scarce
not closely associated in space and time with calc-alkaline igneous rocks	associated in space and time with calc-alkaline granites and volcanics
$SiO_2 \geq 70\%$	$SiO_2 \geq 70\%$
corundum normative	corundum normative
$K_2O \geq Na_2O$	$K_2O \geq Na_2O$
Rb, U, Th, Ta high	Rb, U, Th, Ta high
Sr, Ba low	Sr, Ba low
B high	B low
mineralised, Sn, W, U, Cu, Zn, Pb	not mineralised
extensive pneumatolytic and hydrothermal alteration	very limited pneumatolytic, and hydrothermal alteration

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Table 2. Major element and Normative Data

	Dartmoor					Cairngorm			
	D286	D339	D346	D120	D347	CN48	CN26	CN1b	CN28
SiO_2	73.43	74.87	74.97	75.61	76.09	75.25	76.37	76.82	76.89
TiO_2	0.34	0.22	0.3	0.17	0.11	0.13	0.09	0.09	0.11
Al_2O_3	12.52	12.62	12.67	12.71	12.48	12.81	12.76	12.47	12.32
Fe_2O_3	2.91	2.1	2.52	1.77	1.36	1.2	0.79	0.88	1
MnO	0.09	0.04	0.06	0.09	0.04	0.05	0.04	0.06	0.05
MgO	0.33	0.19	0.35	0.1	0.1	0.03	-	-	-
CaO	0.67	0.42	0.76	0.42	0.34	0.74	0.48	0.32	0.46
Na_2O	2.56	2.88	2.86	3.01	2.96	3.52	3.5	3.5	3.64
K_2O	5.32	5.31	4.97	4.7	5.02	5.04	5.38	4.81	4.85
P_2O_5	0.21	0.16	0.15	0.18	0.08	0.04	0.03	0.01	0.02
LOI	0.6	0.56	0.47	0.5	0.55	0.32	0.23	0.34	0.28
Sum	98.98	99.37	100.08	99.26	99.13	99.13	99.67	99.3	99.62
CIPW Norms:									
Q	35.42	35.37	35.65	37.71	37.22	32.16	32.42	35.37	34.13
C	2.06	1.97	1.75	2.63	1.97	0.34	0.42	1.06	0.33
or	32.68	32.33	30.08	28.63	30.59	30.47	32.29	29.06	29.17
ab	23.9	26.66	26.31	27.87	27.42	32.35	31.92	32.14	33.28
an	2.1	1.07	2.86	0.94	1.2	3.49	2.22	1.56	2.19
hy	1.02	0.58	1	0.45	0.38	0.16	0.04	0.11	0.07
mt	1.9	1.36	1.62	1.14	0.88	0.77	0.5	0.57	0.64
il	0.49	0.32	0.43	0.24	0.16	0.17	0.13	0.11	0.16
ap	0.43	0.34	0.32	0.39	0.17	0.09	0.06	0.02	0.04

Total iron reported as Fe_2O_3. Norms calculated by setting $FeO/Fe_2O_3=0.6$, after Exley and Stone (Appendix B, in Sutherland, 1982)

Table 3. Trace Element Data (ppm)

	Dartmoor					Cairngorm			
	D286	D339	D346	D120	D347	CN48	CN26	CN1b	CN28
Rb	513	555	443	628	564	452	460	639	557
Sr	52	28	54	18	26	68	27	37	25
U	13	17	14	25	14	17	22	18	22
La	42	34	34	18	19	62	52	33	49
Ce	92	67	63	49	38	81	73	45	68
Nd	40	32	28	22	18	73	41	35	35
Sm	8.9	7	6.1	4.6	4.1	11	10	8.9	9.5
Eu	0.69	0.39	0.61	0.34	<0.04	0.19	0.11	0.29	0.15
Tb	1.2	1	0.8	0.6	0.7	1.3	1.3	1.3	1.2
Yb	3.8	3.5	3.1	1.9	3	6.8	8.3	8	6.5
Lu	0.59	0.6	0.47	0.27	0.58	0.84	1	0.88	0.91
Y	46	44	39	32	41	63	76	88	73
Zr	181	129	131	87	86	122	101	102	107
Hf	7.8	5.1	4.3	4.3	3	4.4	3.5	3.5	3.5
Nb	22	18	16	20	13	40	41	65	57
Ta	9.4	11.6	6.4	10.3	7.1	5.3	5	7.4	6.1
Th	29	25	26	15	21	37	33	25	32

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Major element geochemistry

Although many studies of specific aspects of the south-west England granites have been published (see, for example, bibliography for Part 5, Sutherland, 1982), whole-rock data for biotite-granite--the dominant rock in most of the plutons--are not abundant and the intrusions of the Cairngorm group are even more sparsely documented. Our own analyses (Tables 2 and 3), combined with those cited in the literature, show that both batholiths consist largely of corundum-normative rocks with SiO_2 > 70% and K_2O > Na_2O. These rocks are true granites and have a low colour index; MgO + FeO + Fe_2O_3 generally total < 2%.

The compositions of the granites are consistent with the low relative densities implied by the large Bouguer gravity anomalies centred on the two batholiths. Brooks *et al.* (1983) using as a basis the work of A1-Rawi, account for the anomaly over the Cornubian batholith in terms of fixed density contrasts of 0.10 and 0.15 gcm^{-5} to at least 9 km depth under Bodmin and at least 12 km under Land's End. Dimitropoulos (1981) used similar contrasts to explain the anomaly over the Cairngorm batholith. The apparent persistence of the inferred density contrast with the enclosing rocks down to depths of 10 km shows that there is little vertical variation in the major-element composition of the batholiths: a rapid downward passage to less silicic rocks of grandioritic or dioritic character would result in downward reduction of the density contrast between the batholith and its host rocks.

Trace element geochemistry

The distinctive character of the trace element composition of the batholiths is apparent from regional geochemical maps based on the analysis of stream sediments (Webb *et al.,* 1978, B.G.S., in prep.) which show anomalously high values of Be, Li, Rb, U centred on granite outcrops (Watson and Plant, 1979, Johnstone *et al.,* 1979). These generalised results are confirmed and extended by published lithogeochemical data, (see e.g. Simpson *et al.,* 1979, Plant *et al.,* 1980, Sutherland, 1982, Table B.16), and by the authors' own results (Tables 2 and 3). Figure 2, in which our values for 18 petrogenetically important elements are normalised to chondritic abundances displays the distinctive features and emphasises the striking similarity between the signatures of rocks from the Cornubian and Cairngorm granites.

The essential features that characterise the class of granite to which the Cornubian and Cairngorm batholiths belong and distinguish it from more "normal" calc-alkaline varieties are shown by the evidence given in this paper and in Plant *et al.,* 1980, 1983 and Simpson *et al.,* 1979. These differences include:

very high Rb, Th, U, Ta (also Be, Li)
low Nb/Ta, Th/U

very low Ba, Sr, Ti

large negative EU anomalies on chondrite-normalised Ree plots

high Rb/Sr, Rb/Ba, etc.

Notwithstanding the obvious similarities of the two batholiths, there are a few subtle differences, perhaps relating to the details of the fractionation scheme: Dartmoor has a smaller negative P anomaly (restricted apatite removal?) and a more fractionated (i.e. lower) Nb/Ta ratio. Cairngorm has systematically slightly higher HREE and Y abundances and a rather higher Zr/Hf ratio. It is appropriate to note here that these trace element variations are compatible with the effects of extreme fractional crystallisation, provided that the parent melt was of an appropriate type (see later).

Although the extent to which the distinctive trace element signatures persist in depth is uncertain, two groups of observations suggest that vertical variations are small Plant *et al.,* (1980) found no consistent differences in U content between samples collected over a vertical spread of more than a kilometre in the rugged Cairngorm massif. In south-west England, the exceptionally high present-day heat flow is attributed to the high content of the heat-producing radioelements U, Th, K in the granites (Tammemagi and Smith, 1975). Richardson and Oxburgh (1978) infer from their data that these elements must have a rather uniform vertical distribution down to a maximum depth of 16kin and it does not seem unreasonable to suppose that other incompatible elements which have similar geochemical properties show a similar distribution.

Forms of emplacement

Gravity data show that the Cornubian batholith as a whole has steep, somewhat irregular walls and a broad, flattish roof on which the exposed granites form low prominences (Brooks *et al.,* 1983, Figure 10.2). The general character of the gravity low and the outcrop-patterns of granite contacts suggest that the Cairngorm batholith has a similar general form. The long axis of each batholith is strongly discordant with respect to the structural grain of the enclosing rocks, a point which will be taken up in Sections 4 to 6.

Like most large granite bodies, the Cornubian and Cairngorm batholiths are made up of many intrusive units emplaced within and alongside one another. Many of the exposed granites in each area appear to occupy spaces opened by piecemeal or wholesale subsidence of blocks of the envelope rocks (e.g. Exley and Stone, 1982, Oldershaw, 1974). These granites certainly represent high-level intrusions some of which probably reached to within a kilometre or so of the land surface: some, such as the Lochnagar ring complex, may have been surmounted by volcanic piles.

Little is known about the lowermost parts of the batholiths. As was noted above, truly granitic materials with low density and high heat production probably extend down to near the mid-point of the crust (10-15km), while rocks with higher densities and seismic velocities and lower heat production are found at greater depths. Bott *et al.,* (1970) inferred from their data that these lower rocks occupied the sites from which the batholith magmas had been extracted by partial melting and that they were made up of the residual lower crustal material together with stoped blocks displaced during the ascent of the magma. Preliminary results of the SWAT seismic traverse reveal a flat layering which seems incompatible with this inference (D. J. Blundell, *pers.*

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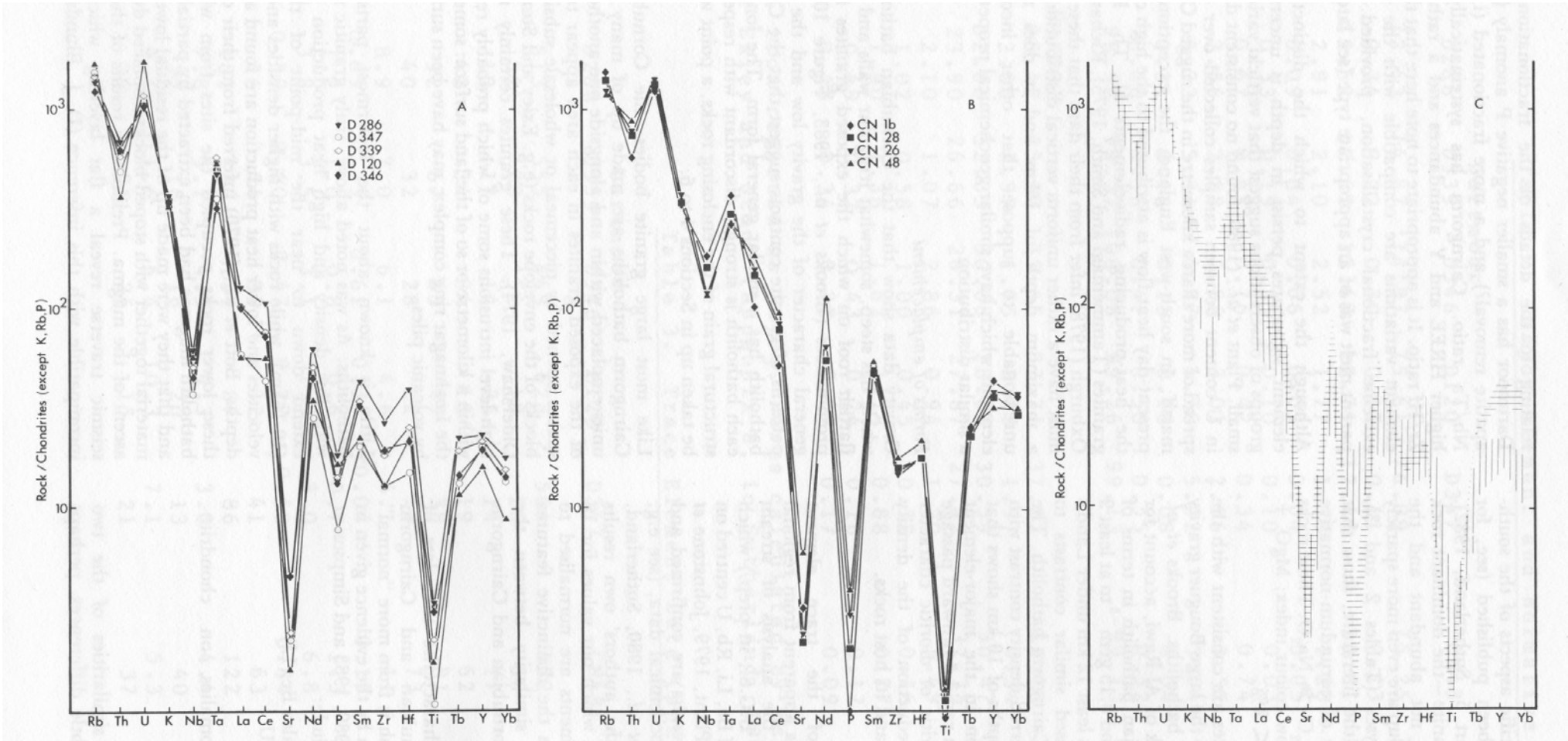

Figure 2. Element abundances normalised to chondrite values for (A) biotite granite samples, Dartmoor (B) biotite granite, Cairngorm (C) Fields of Dartmoor and Cairngorm superimposed. The construction of the diagrams is based on reasoning outlined by Thompson (1982) with the addition of U, using a normalisation factor of 0.013.

comm.). Shackleton *et al.,* (1982), on the other hand, suggest that the batholith has no autochthonous root and that it was fed by magma derived from a southerly source (see Section 5). In the Cairngorm area, the distinctive granitic material forming the batholith in the upper crust also seems to terminate at depths of 10-15kin (Dimitropoulos, 1981) and the only information available as to the nature of the underlying rocks suggests that they have high seismic velocities (Bamford *et al.,* 1977).

4 Crustal setting: the Grampian Highlands

The metamorphic Caledonides of Scotland represent the marginal parts of the former North Atlantic continent which were intensely deformed in Lower Palaeozoic times by collision with one or several continental fragments during the closing of the Iapetus Ocean. The tectonic and thermal evolution of the Grampian Highlands (the host region of the Cairngorm batholith) has been examined elsewhere (Plant *et al.,* 1983, Watson, 1984) and is therefore not discussed in detail: the salient points relevant to the genesis of the batholith are outlined below and in Figures 3 and 4.

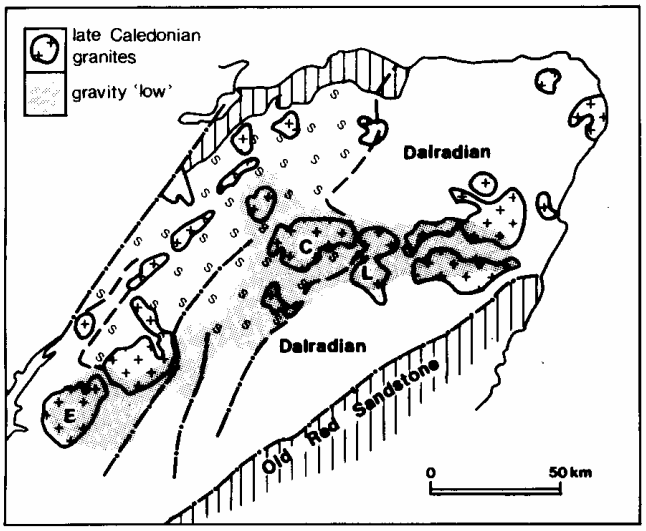

Figure 3. The distribution of late Caledonian granites in the Grampian Highlands of Scotland. The outline of the negative gravity anomaly corresponds to the − 30 isogal and shows the approximate extent of the Cairngorm batholith in the north-east and a possibly buried granite towards the south-west; Most of the smaller granites outside the limits of the gravity anomaly are talc-alkaline. C = Cairngorm, L = Lochnagar, E = Cruachan (Etive) granite.

The crust from which the Grampian Highlands were developed was of continental type and consists of an ancient, highly-metamorphosed basement (Lewisian gneisses and granulites, Moine metasediments) together with a thick Dalradian sedimentary and volcanic sequence deposited only shortly before the beginning of the Caledonian collision period. During this period, roughly 500-430 Ma, oceanic crust from marginal basins or the Iapetus Ocean itself was probably subducted beneath the Highland area and small continental blocks impinged on the area. The Dalradian prism was strongly compressed, together with its basement, and correspondingly thickened to form a NE-SW crustal welt 60-70 km in thickness. As a result of these tectonic events, crustal temperatures rose, regional metamorphism affected the whole area and isostatic uplift set in train a phase of deep erosion. By about 400 Ma, some 25-30 km. of overburden had been stripped from the Highlands by erosion and the crust had once more been reduced to about 30 km thickness, virtually the whole of which was made up of high-grade or medium-grade metamorphic rocks.

The emplacement of late-orogenic granites did not begin in earnest until about 435 Ma, by which time compressional deformation had almost ceased and crustal temperatures (at the levels now exposed) had fallen well below those necessary for metamorphism (Fig. 4). The regional tectonic environment by this time appears to have been dominated by sinistral transcurrent movements parallel to the Caledonian belt on the Great Glen fault and its associates and by limited extension normal to the belt.

The majority of the granites emplaced during the period 435-410 Ma are calc-alkaline and granodioritic. These rocks lack the anomalous geochemical signature of the Cairngorm group and many appear, from trace element and isotope studies, to have been derived largely from sources within the Highland crust. The latest Caledonian granites, emplaced during the period 415-400 Ma, include the Cairngorm batholith, the ring complexes of Cruachan, Glencoe and Ben Nevis (S.W. Highlands), together with some smaller plutons and a variety of porphyrite, lamprophyre and felsite minor intrusions. These bodies reached high crustal levels, largely by cauldron subsidence, stoping and similar mechanisms, and some were associated with surface volcanicity. Most of the latest intrusions have the distinctive geochemical signature illustrated in Figure 2, although a few resemble the earlier members of the late orogenic suite. Contributions from both deep crustal and enriched mantle sources for the parent magmas have been envisaged (e.g. Simpson *et al.,* 1979, Plant *et al.,* 1980, Clayburn *et al.,* 1985).

5 Crustal setting: south-west England

The Variscan province of south-west England and Southern Ireland was developed within or near the southern margin of the north European continental plate. The crust (with the exception of the ophiolitic Lizard complex) is of continental type and has a present thickness of just under 30 km (Bott *et al.,* 1970). The exposed country rocks are volcanic and sedimentary formations of Upper Palaeozoic age, probably forming a succession not much more than 5 km in original thickness. The present thickness has been increased by folding (see later) and is probably of the order of 10-15 km (see generalised cross-sections of Shackleton *et al.,* (!982), Hobson and Sanderson (1983)). The age and nature of the underlying basement are not known, but seismic velocities of 6.0-6.3 kms-1 at depths of less than 8 km beneath the Bristol Channel (Brooks *et al.,* 1983) suggest that metamorphic or igneous rocks may

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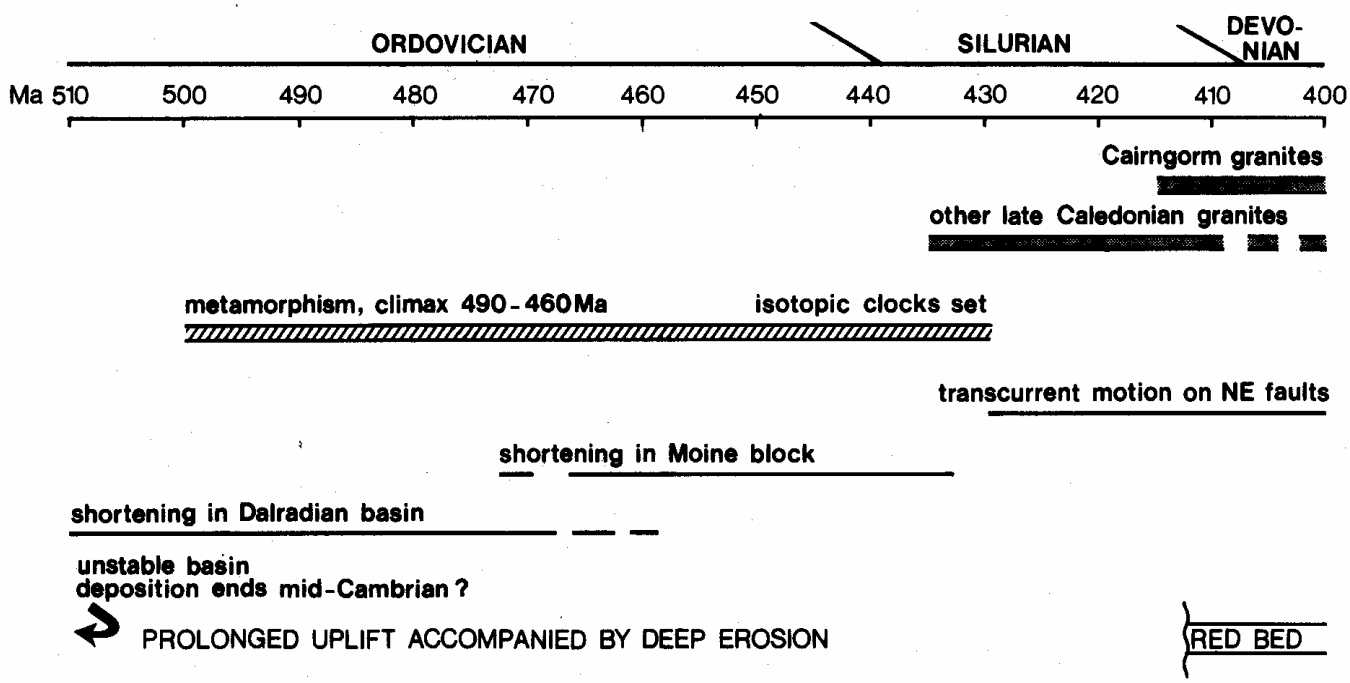

Figure 4. Chart showing the time-relationships of sedimentation, deformation and granite emplacement in the Grampian Highlands.

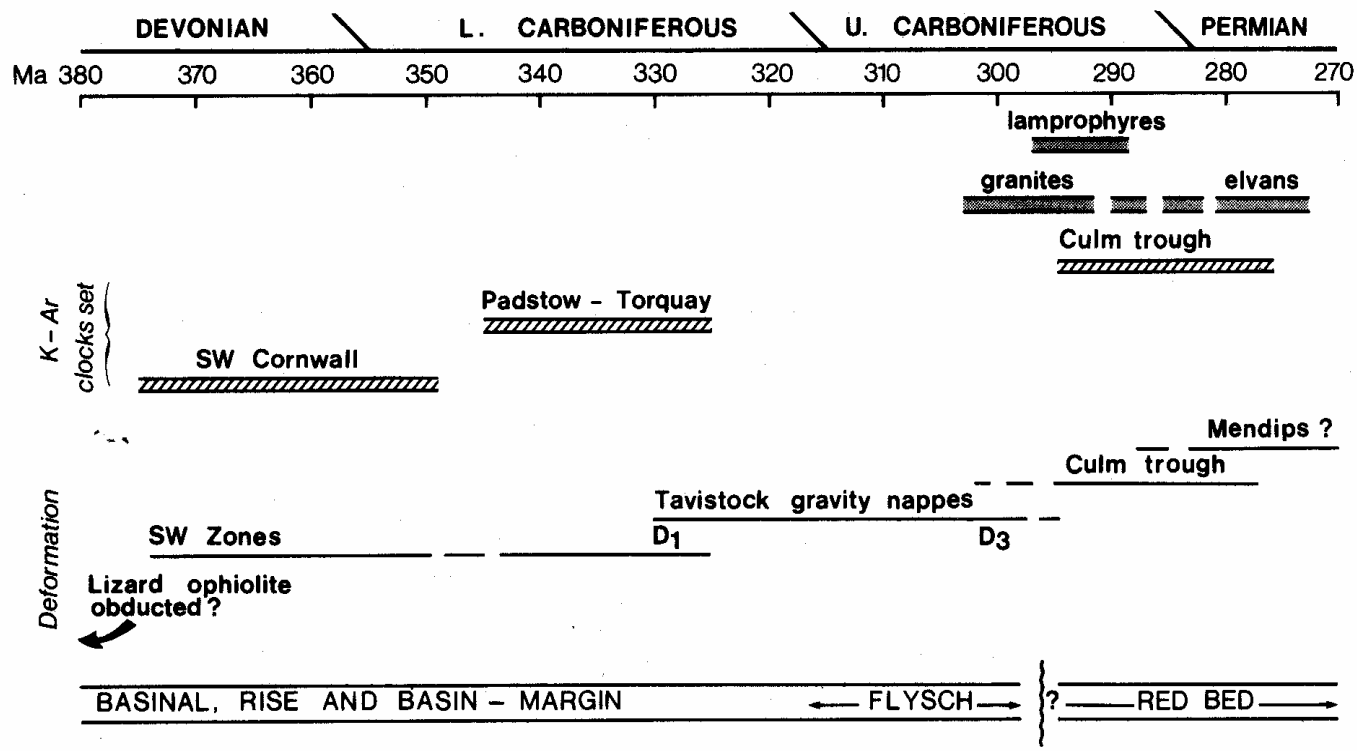

Figure 5. Chart showing the time-relationships of sedimentation, deformation and granite emplacement in south-west England.

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predominate. The basement may be similar to the Precambrian of Normandy and the Channel Islands which includes both a deformed and low-grade Brioverian succession, with Cadomian granites, and older gneisses; or it may consist entirely of late Precambrian rocks (cf Hampton and Taylor, 1983).

The tectonic evolution of the Upper Palaeozoic assemblage in south-west England spanned a period of roughly a hundred million years from early Devonian to late Carboniferous or early Permian (Fig. 5). The intensity of deformation, the reduplication of strata on numerous southward-dipping thrusts and the occurrence of the Lizard obducted ophiolite complex suggest that the region was subjected to tectonic shortening normal to the Variscan front. An interpretation in terms of the elimination of an oceanic basin and consequent collision with one or more continental blocks advancing from the south has been advocated by some authors. There is, however, no clear evidence that northward subduction of oceanic crust beneath the area took place and some authorities doubt whether collision processes were important in the evolution of the province (e.g. Krebs, 1977).

The fact that marine sediments continued to accumulate almost without a break through Devonian and Carboniferous times shows that the crust did not undergo thickening sufficient to produce regional uplift during the main period of tectonic activity. Although the sea was finally expelled from the region late in the Carboniferous period, the subsequent erosion does not seem to have bitten deeply into the deformed rock pile. The uppermost stratigraphical units (the Culm) have a style of deformation and absence of metamorphic effects which suggest that they were not covered by more than a few kilometres of overburden at any stage. The Devonian rocks commonly show effects of ductile deformation and low-grade metamorphism, but there seems to be no evidence that this metamorphism was of high pressure type, An overburden of at most 10-15 km at the climax of metamorphism seems not unreasonable.

The lack of great regional uplift during the period 380-300 Ma implies that the crust had not been greatly thickened, since the development of a low-density crustal welt normally leads to uplift due to isostatic forces. Shackleton *et al.*, (1982) suggest that tectonic shortening (which they estimate at 150 km across a belt now reduced to about 150 km) was confined to the Upper Palaeozoic succession and that the Upper Palaeozoic rocks have been detached from their basement and pushed northward on a low-angle dislocation. This interpretation eases the problem of uplift, because the increase of thickness produced by a reduction to half its width of a layer originally, say 5 km thick would be only about 5 km, insufficient to bring about massive uplift.

The dating of tectonic events in south-west England depends in part on stratigraphic evidence and in part on isotopic dating of metamorphic minerals. In south-west Cornwall, K-Ar mineral ages from Devonian phyllites indicate that the low-grade metamorphism associated with deformation had been completed and regional temperatures had fallen below the level (300-350°C) at which the isotopic "clocks" were set by about 350 Ma (cf Hawkes 1981). In central Devon, gravitational nappes piled up and were refolded in early Namurian times (Isaac *et al.*, 1982) and folding in the Culm basin continued into late Westphalian times. Taking south-west England as a whole, the period of severe deformation lasted roughly a hundred million years (Fig. 5).

The emplacement of the Cornubian batholith came towards the end of this period and the exposed granites are generally regarded as post-tectonic (Mitchell, 1974). When considered on a local scale, the siting of these granites reveals the influence of the local structure--in particular, the Dartmoor and Bodmin granites are located at the southern margin of the main Culm basin where a submarine (perhaps fault-controlled) ridge persisted through Visean and early Namurian times (Isaac *et al.,* 1982, Fig. 3). On the grand scale, however, the tong axis of the batholith as defined by the negative gravity anomaly is strongly discordant not only with respect to stratigraphical boundaries and structural grain but also with respect to the broad zones defined by the style and age of their structures (cf Sanderson and Dearman, 1973).

Two possible explanations for the large-scale discordance of the batholith are worth considering. A shift in the regional stress system at the end of the long period of shortening on N-S lines could have led to the opening of oblique fractures allowing the ascent of granite magma (cf Hawkes, 1981). Normal faulting in central Devon gives evidence that localised north-south extension took place in early Westphalian times (Isaac *et al.,* 1982) and the alignment of dykes and mineral veins implies that there was extension on similar lines after the rise of the batholith (Moore, 1975). It will be suggested that these movements reflect a general change in the controlling tectonic regime (see Section 7).

Alternatively, the discordant trend of the batholith with respect to the structure of the Upper Palaeozoic may have been determined by structures in the underlying basement or upper mantle. The very low metamorphic grade of the Upper Palaeozoic and, still more, the isotopic evidence showing that in some parts of the region crustal temperatures had begun to decline fifty million years before the rise of the granites, make it unlikely that conditions suited to large-scale partial melting existed at mid-crustal levels. To attain temperatures of 7500-800° near the known base of the batholith, one must postulate an unrealistic geothermal gradient of about 50°C km-1. The sites of magma formation were therefore probably much deeper, within or below the pre-Variscan basement. Shackleton *et al.*, (1982) suggest that these sites lay well to the south of the batholith and that magma moved laterally to its present location along a south-dipping detachment zone at the base of the Upper Palaeozoic (see earlier). It seems better, however, to seek for the sites of magma formation directly below the batholith, since the buoyancy imparted by the low density of the magma would favour vertical rather than subhorizontal flow. We therefore suggest that the magma followed a major WSW-ENE basement dislocation parallel to the Ouessant-Alderney line of disturbance on the floor of the English Channel which is regarded as a

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rejuvenated basement dislocation, possibly an early Palaeozoic suture (Lefort, 1977). Similar dislocations, possibly controlled by basement structure are recorded west of the Bristol Channel.

6 Caledonian-- Variscan comparisons

The details summarised in Sections 4 and 5 provide the basis for Table 4 in which various aspects of the crustal settings of the batholiths are compared. As regards the nature of the crust itself, the principal common feature is the predominance in the lower portion of the crust of rocks which were already crystalline and largely dehydrated before the period of magma formation. In other respects, the host provinces have little in common.

So far as tectonic history is concerned, both provinces lay near active plate margins and had been subjected to a long period of compressional deformation prior to the formation of the batholiths. The intensity of regional metamorphism and the extent of regional uplift and consequent unroofing were very different in the two provinces.

The most significant points of resemblance seem to be those connected with events actually during the period of batholith formation. In both provinces, the magma rose at the end of a long period of penetrative deformation. The effects of Orogenic stresses seem to have died away, at least in the vicinity of the batholiths, where limited extension took place. Block movements on dislocations extending to the base of the crust are clearly documented in the Grampian province. The discordant habit of the Cornubian batholith and the occurrence in the English Channel and Western Approaches of major dislocations parallel to the batholith axis suggest that granite magma gained access to the upper crust by means of deep fractures. Geophysical indications that both batholiths

TABLE 4. THE GEOLOGICAL SETTING

South-west England	**Grampian Highlands**
The Crust	
Type and present thickness	
Continental, 25-30 km	Continental, 25-30 km
Tectonic province	
Variscan, stabilised c 280 Ma	Caledonian, stabilised c 400 Ma
Lower crust	
Moderate seismic velocity, probably not depleted in LIL elements, may be Late Proterozoic as in Brittany and Channel Islands	High seismic velocity, depleted in LIL elements, largely Archaean to mid-Proterozoic gneisses and granulites
Middle and upper crust	
Upper Palaeozoic sediments and volcanics, low metamorphic grade	Late Proterozoic to Cambrian metasediments, usually high metamorphic grade
Tectonic developments prior to batholith emplacement	
Tectonic shortening without marked thickening of crust: intermittent deformation lasting c 100 m.y.	Severe tectonic shortening and marked thickening of crust to c 60km: intermittent deformation lasting c 100 m.y.
Regional uplift begins late, limited in extent	Regional uplift prolonged
No deep erosion	Deep erosion, 25-30 km of overburden removed
Metamorphism low-grade, climax precedes batholith by up to 50 m.y.	Metamorphism generally medium to high grade, climax precedes batholith by ≥ 40 m.y.
? No subducted oceanic crust descending beneath area	Subducted oceanic crust descending beneath area prior to batholith emplacement.
Emplacement of batholith	
Followed period of tectonic shortening?	Followed period of tectonic shortening
Local extension normal to orogenic belt at about time of emplacement	Transcurrent fault movements and local extension at about time of emplacement
Axis of batholith discordant to structural grain in upper crust	Axis of batholith discordant to structural grain in upper crust
Magma originates near base of crust, rises to near-surface levels	Magma originates near base of crust, rises to near-surface levels
Generation and ascent of magma facilitated by block movements?	Generation and ascent of magma facilitated by block movements
Rise of magma followed by wide-spread pneumatolytic and hydrothermal activity	Pnuematolytic and hydrothermal activity very limited

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pass at 10-15 km depth into rocks with higher densities suggest that little or no intrusive material was retained in the lower crust.

7 Petrogenetic Implications

The details summarised above lend no support to the idea that the geochemical resemblances between the Cornubian and Cairngorm batholiths are the result of derivation from similar parent materials within the crust itself. The deeper parts of the crust (which alone might have yielded large volumes of magma) have different characteristics in south-west England and the Grampians, (Table 4). From the geophysical evidence, it seems very unlikely that a large mass of wet sediment could have been incorporated in the lower crust of either province at the time of formation of the batholiths (c.f. Plant *et al.*, 1984) and we therefore do not follow Halliday (1984) in his proposal that such material was the source of the Grampian batholith.

Some of the isotopic and trace element characteristics of the granites have, nevertheless, been held to imply a crustal origin (e.g., Hampton and Taylor, 1983); more specifically, a source in deeply buried wet sediments has been inferred from the resemblance of the batholiths to the S-type granites of Chappell and White (1974). The interpretation of the data is, however, controversial. We have already put forward reasons for concluding that the 'package' of S-type characteristics can be more satisfactorily explained in terms of interactions between a magma of sub-crustal origin and the crust through which it rises (Plant *et al.*, 1980). Some aspects of the geochemical signature seem incompatible with a wholly crustal derivation. The partitioning of trace elements during partial melting depends upon their compatibility with respect to mineral phases which remain stable in the solid residuum. The very high levels or uranium, thorium and tantalum in the granites (Fig. 2) suggest that the melt was derived from sources lacking in stable minerals with an affinity for these elements. Deep crustal rocks commonly contain accessories such as zircon, sphene, allanite and ilmenite Which can accommodate U, Th, Ta, whereas such accessories are rare in mantle rocks. More specifically, the high levels of large ion lithophile (LIL) and high field strength (HFS) elements in the Cairngorm batholith are conspicuously anomalous with respect to the Moine, Dalradian and Lewisian country rocks that make up the enclosing crustal column (Johnstone *et al.*, 1979). The Moine and Dalradian assemblages differ between themselves with respect to key elements such as yttrium and zirconium, but the Cairngorm granites which transgress the Moine-Dalradian boundary maintain almost constant levels of Y and Zr. These discrepancies lead us to conclude that the parent magmas of the batholiths originated below rather than in the crust.

The search for common features in the geological setting which could account for the resemblances between the Cornubian and Cairngorm batholiths leads to the idea that the crucial factors were those connected with the tectonic regime in operation at the time of magmatism. In both provinces, a long phase of orogenic compression gave way to a period of limited extension accompanied by block movements on deep faults. We follow Leake (1978) in supposing that the relief of pressure and disturbance of thermal equilibrium resulting from block movements were instrumental in creating pressure-temperature conditions favourable to partial melting at or below the base of the crust. The role of regional uplift, which has been emphasised by some authorities (e.g. Pitcher, 1982) is thought to have been of secondary importance, because the scale and timing of uplift were very different in the two provinces.

The most probable source for the parent magmas appears to be mantle material previously enriched in volatile and incompatible elements by metasomatic processes (Simpson *et al.*, 1979). The distinctive geochemical signatures of the two batholiths and the family resemblances to Variscan tin granites elsewhere in Europe point to common features in the processes of partial melting and subsequent magmatic evolution (cf Lehmann, 1982). The details of the multi-element diagrams with their pronounced peaks and troughs (Fig. 2) are compatible with extensive fractional crystallisation before and during the ascent of the magma. St, P, Zr and Ti, all showing moderate or low abundances, may have been abstracted during the crystallisation of such minerals as plagioclase, apatite, zircon and sphene while the more abundant LIL and HFS elements, U, Th, Rb, Ta were retained until the final stages of consolidation. The huge bulk of granite material in the upper crust must be balanced somewhere at depth by a dense, basic residuum left after extraction of the partial melt. There seems no indication that this residuum is contained within the crust and we suggest that it remained at or beneath the moho, where the magma was generated (cf Simpson *et al.*, 1979).

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Read at the Annual Conference of the Ussher Society, January 1984

Sequence of coralline faunas and depositional environments in the Middle Devonian Daddyhole Limestone Formation stratotype section, Torquay, Devon.

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Goodger, K. B., Buglass, A. and Scrutton, C. T. 1984. Sequence of coralline faunas and depositional environments in the Middle Devonian Daddyhole Limestone Formation stratotype section, Torquay, Devon. *Proceedings of the Ussher Society, 6,* 13-24.

A correlation is proposed for the major outcrops in the stratotype section of the Middle Devonian Daddyhole Limestone Formation, on the south side of the Torquay promontory, South Devon. The coralline faunas in the approximately 125 m thick sequence are analysed and the petrography of the limestones outlined. Several tabulate and rugose coral associations are recognised, which appear to be characteristic of different environmental conditions. The succession is interpreted as representing the migration into the area of the edge of a carbonate platform which built up rapidly to intertidal level. Environments in the lower third of the sequence were largely somewhat restricted. A wedge of shales, the Knoll Shale, is present on the platform margin near the top of these facies. Above, stromatoporoids colonised and eventually dominated the local platform margin in less turbid, fully marine conditions. This probable bioherm is buried on its flanks by more rapidly accumulating sediment in which a rich fauna of rugose and tabulate corals is preserved. Towards the top of the limestones a return to more restricted conditions, probably with continuing high rates of sedimentation, completely excluded coralline faunas.

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Introduction

In the Devonian of south-west England, extensive carbonate platforms are developed within the dominantly marine sequences of South Devon. The carbonate platform successions which crop out inland from Tor Bay show contrasting marginal and internal facies of Givetian age interpreted as the result of reef build-ups fringing at least the eastern margins of the platform. This development has been described as the Tor Bay Reef-Complex by Scrutton (1977a, b). On the platform margin in the Torquay area, the reef facies of the Walls Hill Limestone Formation is underlain by the lowest of the three formations in the Torquay Limestone Group, the Daddyhole Limestone (Fig. 1). This limestone is a sequence of dark to medium-grey, predominantly well-bedded carbonates with some minor shaley interbeds and subsidiary shale members, representing the initiation and pre-reefal early development of the platform margin. Much of the sequence is richly fossiliferous with corals, stromatoporoids, bryozoa, brachiopods and crinoids as ossicles most prominent, but in addition with bivalves, gastropods, some trilobites and a varied microfauna. As a whole, the fauna indicates an early to late Eifelian age for the Daddyhole Limestone (Scrutton 1977b, 1978).

This limestone crops out in several localities across the southern half of the Torquay peninsula and correlation between the structurally isolated sections is tentative or uncertain (Scrutton 1977b). Our purpose here is to outline the distribution of the most prominent faunal elements, the coralline assemblages, in the type section in the Daddyhole block (Fig. 2). Even within this mass, cropping out between Triangle Point (92866281) and Torquay Harbour (918651), the lack of reliable market horizons combined with the folded and faulted nature of the section (Scrutton 1978, Fig. 12d) and the inaccessibility of some cliff faces limits the detection and documentation of lateral facies variations. A tentative correlation of the stratigraphic logs from the main exposures in this block is presented in Fig. 3. The limestone here is approximately 125 m thick.

A summary of these successions, their petrography and correlation is followed by an account of the coralline faunas and finally an interpretation of the environments of deposition. K.B.G. and C.T.S. are responsible for the identification and discussion of the coralline faunas, particularly the tabulate corals and the rugose corals respectively. A.B. has provided the petrographic information (see also Buglass 1982) and A.B. and K.B.G. have measured the logs. Grid references all refer to the SX 100 km x 100 km square of the National Grid.

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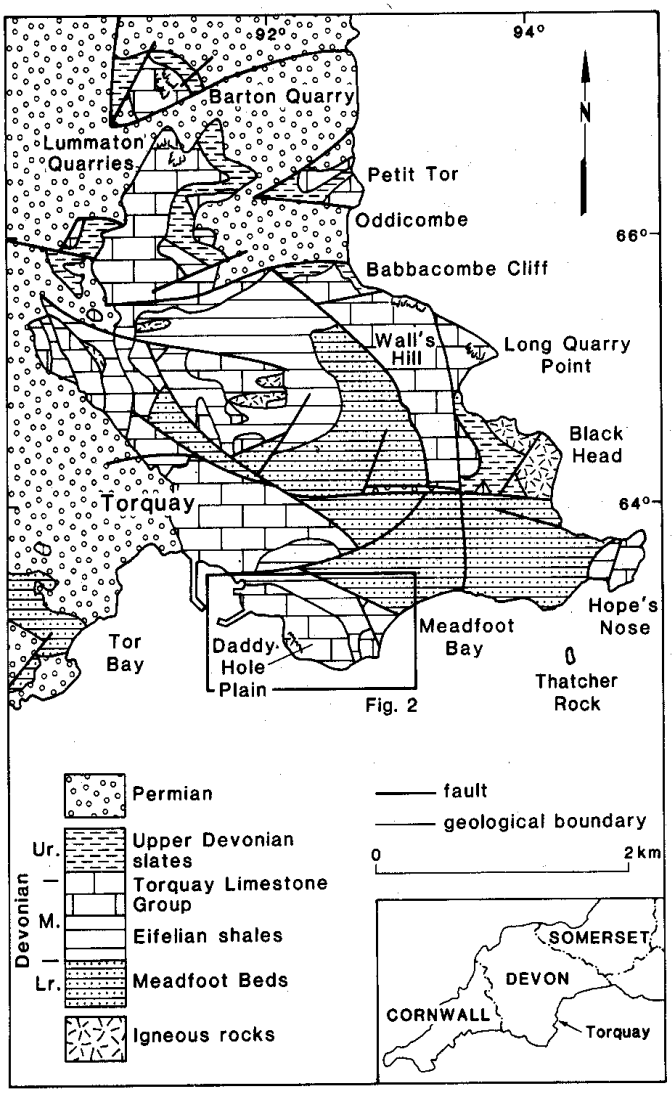


Figure 1. Geological map of Torquay. The area enlarged in Figure 2 is indicated (after Scrutton 1978, Fig. 10b).

Locality details

Daddyhole Cove (927628)

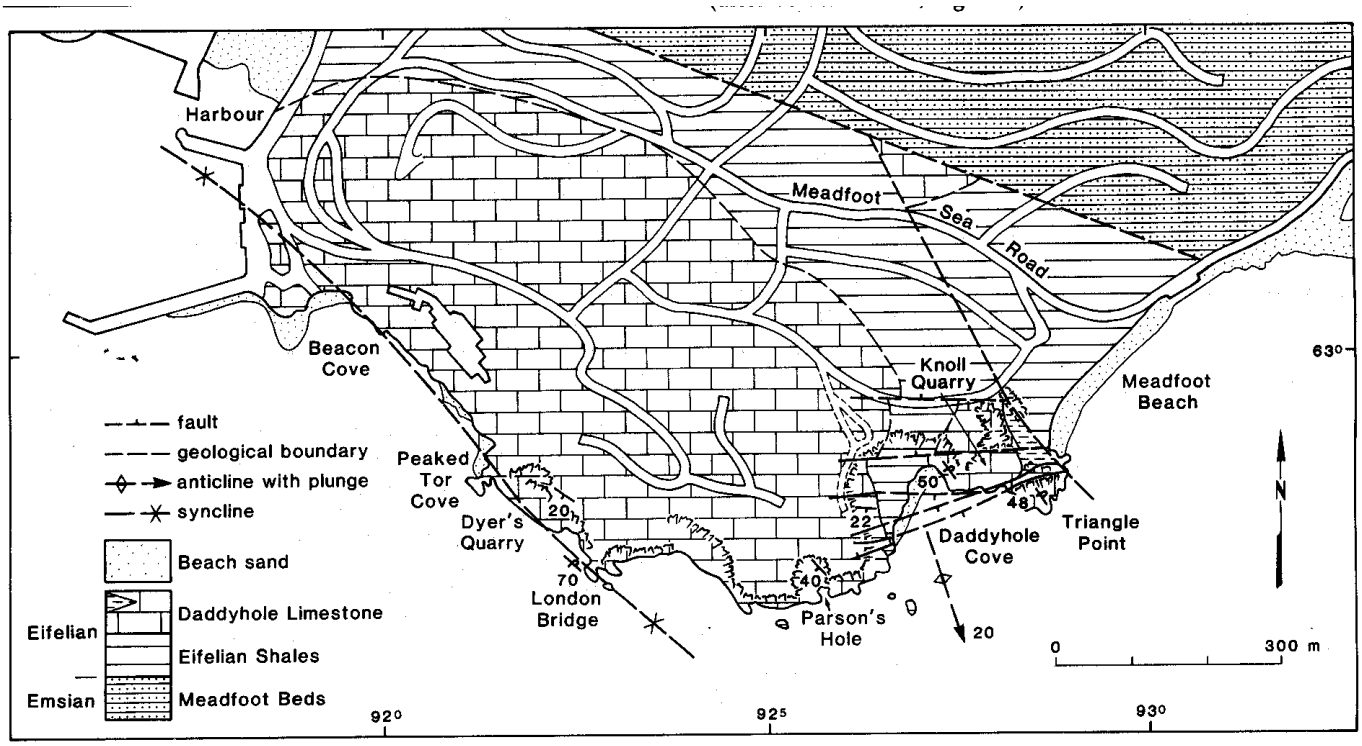
The base of the Daddyhole Limestone Formation is defined at the base of the 5 m sequence of 50-150 mm thick limestone-shale interbeds on the west side of Daddyhole Cove (92666277) (Scrutton 1977b, p. 167; Figs 2 and 3). The limestones are coarse crinoidal packstones. Prominent bryozoan debris and scattered corals, most notably *Calceola sandalina* (Linné) which is frequently oriented flat-face downwards in presumed growth-position (Champernowne, 1874), occur mainly in the shales. Well-bedded crinoidal grainstones succeed the transitional beds higher in the cliff.

The limestone overlies the "Eifelian shales" in the core of a recumbent anticline overturned to the north-east (Fig. 2). Daddyhole Cove is eroded in these less resistant shales. On the overturned limb of the fold, the lower 45 m of the Daddyhole Limestone can be best examined in the Knoll Quarry, or more cleanly weathered, in the old quarry on Triangle Point.

Triangle Point (92866281)

Triangle Point is faulted against the Knoll and at its western end, the 5 m basal sequence of limestone-shale interbeds has been removed by erosion. Detailed logs show that although the two successions are very similar, specific biostromes on Triangle Point cannot be matched

Figure 2. Geological map of the Daddyhole block, Torquay (after Scrutton 1978, Fig. 12a).



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precisely in the Knoll Quarry even over this short distance. Such lateral facies variation is supported by the impersistence of some biostromes along strike on the point.

The sequence of limestones on Triangle Point shows a broad, approximately equal threefold division into basal crinoidal grainstones, succeeded by peloidal grainstones and followed by peloidal and bioclastic packstones (Fig. 3). Biostromes are distributed throughout this succession. Six in the middle and upper parts of the crinoidal grainstones are *Thamnopora* dominated with sparse Rugosa and, at some horizons, tabular and domal stromatoporoids. The peloidal grainstones are notably lacking in coralline faunas except for abundant auloporid fragments in the top 2 m. The grainstones contain common thin graded beds, with fine crinoid ossicles conspicuous at their bases. Possible charophyte oogonia occur around 23 m above the base of the section. At the base of the peloidal packstones, two prominent biostromes, both I m thick, occur 0.70 m apart. The overlying 2.5 m of packstones, with fenestrae and small burrows, contain high-spired gastropods, ostracods and nodular and encrusting algae: there are two prominent levels with desiccation cracks. In the succeeding bioclastic packstones, debris and sometimes whole or partial colonies of the auloporid *Remesia* are common. One of several thinner biostromes forms a prominent bedding plane at the east end of the point, on which globular stromatoporoid coenostea up to 600 mm diameter, many with intergrown *Syringopora*, are scattered. At the top of this sequence, alternations of peloidal and bioclastic packstones with thin shale partings represent the beginning of a transition to shales, better seen on the Knoll.

Interbedded peloidal and bioclastic packstones of similar type crop out at the base of the Parson's Hole succession and also on Saddle Rock (92136282, off Peaked Tor Cove) and in Beacon Cove (91906305).

The Knoll (928629)
Above a sequence of interbedded bioclastic packstones and reddened peloidal packstones, a 3 m unit of fine bivalve bearing packstone is succeeded by 25 m of dark grey shales, sparsely fossiliferous with crinoid ossicles, brachiopods and bryozoa, forming the slope above Triangle Point. There is a 200 mm tuff 20 m above the base of the shale. This shale unit is here designated the Knoll Shale Member. On the western, upright limb of the Daddyhole anticline, there is no sign of the Knoll Shale. The geometry of the fold suggests that equivalent beds form the floor and/or western wall of the disused quarry known as Parson's Hole. It is difficult to allow for differences in rates of deposition and compaction ratios between the equivalent mudrocks and the limestones. However, a 200 mm tuff is present some 9 m above the base of the sequence on the western side of Parson's Hole which we tentatively correlate with that in the Knoll Shale.

Parson's Hole (92556270)
Interbedded bioclastic and peloidal packstones with prominent *Remesia*, similar to those at the top of the Triangle Point sequence, form the lower 10 m of the succession on the west side of Parson's Hole. High-spired gastropods are prominent in some of these beds. Thin shaley partings occur below the tuff and the upper 5 m includes unbroken *Remesia* colonies on some bedding planes and levels rich in *Thamnopora* and *Scoliopora* and scattered stromatoporoids, some encrusting the tabulate corals. Above a 2 m unit of grainstones is a less well-bedded, thick sequence, accessible only in part, which appears to be dominated by bindstones, with *Alveolites* and then *Planocoenites* prominent at lower horizons, joined above by tabular stromatoporoids. The latter are of increasing importance upwards associated with *Alveolites*. They reach at least 2 m diameter with a reddened, dolomitised, bioclastic wackestone/packstone matrix. Present outcrop does not allow the possibility that this is a local stromatoporoid bioherm to be tested.

Minor faulting and shearing complicate thickness measurements in the mid part of the Parson's Hole Sequence.

Dyer's Quarry (92236277)
Dyer's Quarry, the next extensive, accessible outcrop to the west, exposes the major complimentary syncline to the Daddyhole anticline. There are no key horizons allowing a secure correlation with Parson's Hole. Our tentative correlation is based on three factors, including the geometry of the cliff profile as seen from offshore (Scrutton 1978, Fig. 12d). In addition, the Parson's Hole sequence lacks any sign of the richly fossiliferous *Thamnophyllum* bafflestone, prominent in Dyer's Quarry and known to have some lateral extent from outcrops on Thatcher Rock (944628). Finally, the core of the syncline at the western end of Dyer's Quarry displays a sequence thinner but similar to that in the upper part of Parson's Hole. Above coarse crinoidal packstones, *Planocoenites-Alveolites* bindstones rich in *Cladopora* are succeeded by some 5 m of *stromatoporoid-Alveolites* bindstones forming the lower 9 m of the Dyer's Quarry sequence. The correlation proposed between these levels envisages Dyer's Quarry as situated on the flanks of the much thicker stromatoporoid bindstone facies of Parson's Hole. Thus we regard the sequence in the seaward cliff, quarry floor and back wall in the upright limb of the Dyer's Quarry syncline to be wholly younger than the Parson's Hole succession.

This succession in the upright limb is notable for the lower 5-6 m of interbedded *Planocoenites* bindstone, *Thamnophyllum* bafflestone and rudstones--floatstones of solitary rugose corals with a bioclastic wackestone to packstone matrix (Fig. 4f). The latter, dominated by *Lekanophyllum cylindricum* (Schluter) and *Stnngophyllum buechelense* (Schluter), are now almost all recumbent although a few shorter coralla at quarry floor level still stand upright in growth position. On the quarry floor itself, recumbent straight coralla, almost exclusively of *MesophyUum laeve* (Schulz), up to 600 mm (in one doubtful case up to 1.1 m) long show clear alignment (Figs. 4b, 5). The back wall of the quarry consists of microbioclastic packstones, with scattered corals in the lower part and with some 3 m of *Planocoenites* bindstone just beneath a prominent tuff. The packstones above the tuff contain gastropods and thin bivalve shells, but appear to lack a coralline fauna and are poorly accessible. They represent the youngest limestones cropping out in the Daddyhole block.

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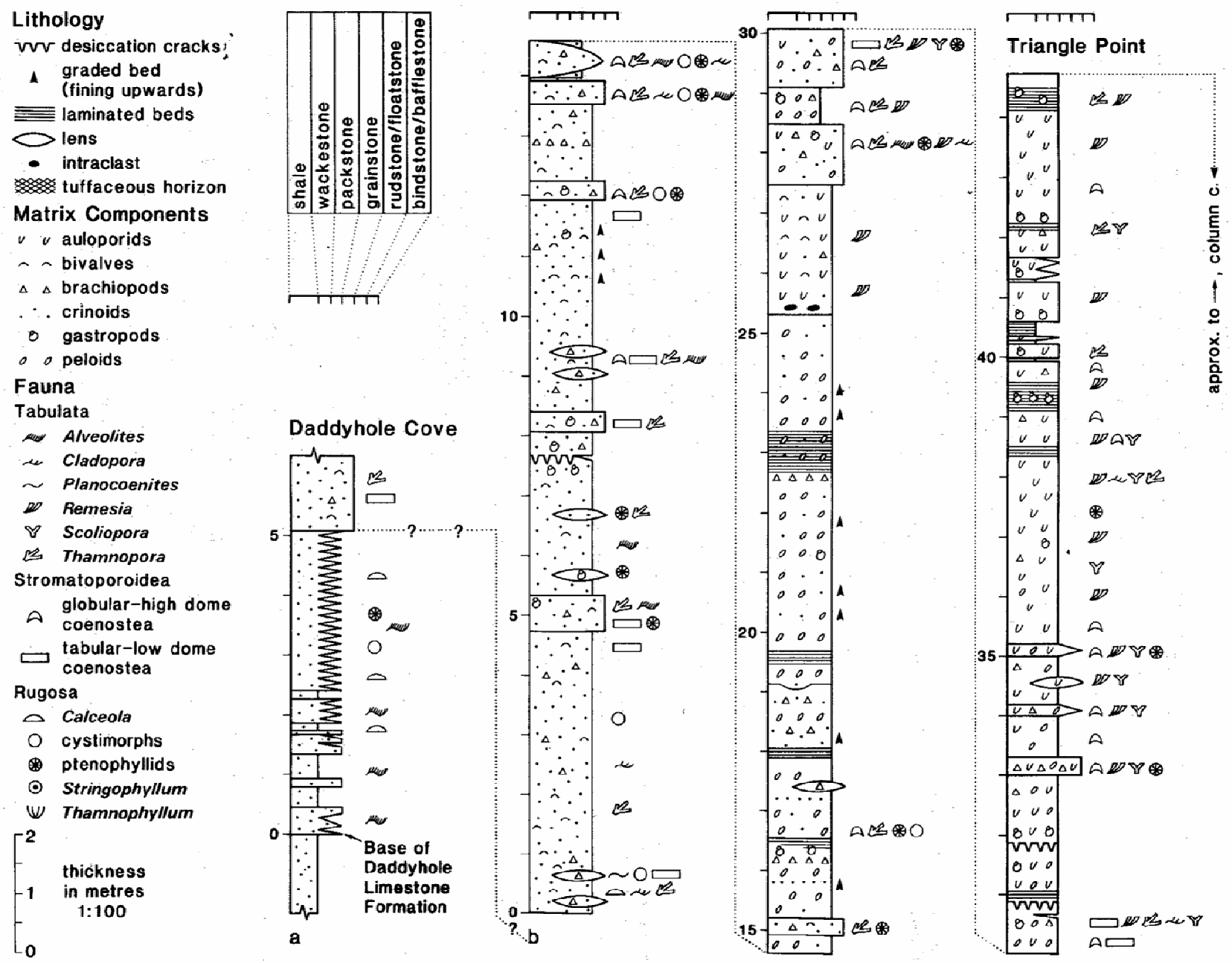
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desiccation cracks
graded bed (fining upwards)
laminated beds
lens
intraclast
tuffaceous horizon
Matrix Components
auloporids
bivalves
brachiopods
crinoids
gastropods
peloids
Fauna
Tabulata
Alveolites
Cladopora
Planocoenites
Remesia
Scoliopora
Thamnopora
Stromatoporoidea
globular-high dome coenostea
tabular-low dome coenostea
Rugosa
Calceola
cystimorphs
ptenophyllids
Stringophyllum
Thamnophyllum
thickness in metres 1:100
shale
wackestone
packstone
grainstone
rudstone/floatstone
bindstone/bafflestone
Daddyhole Cove
Base of Daddyhole Limestone Formation
a
b
Triangle Point
approx. to ⟶, column c.

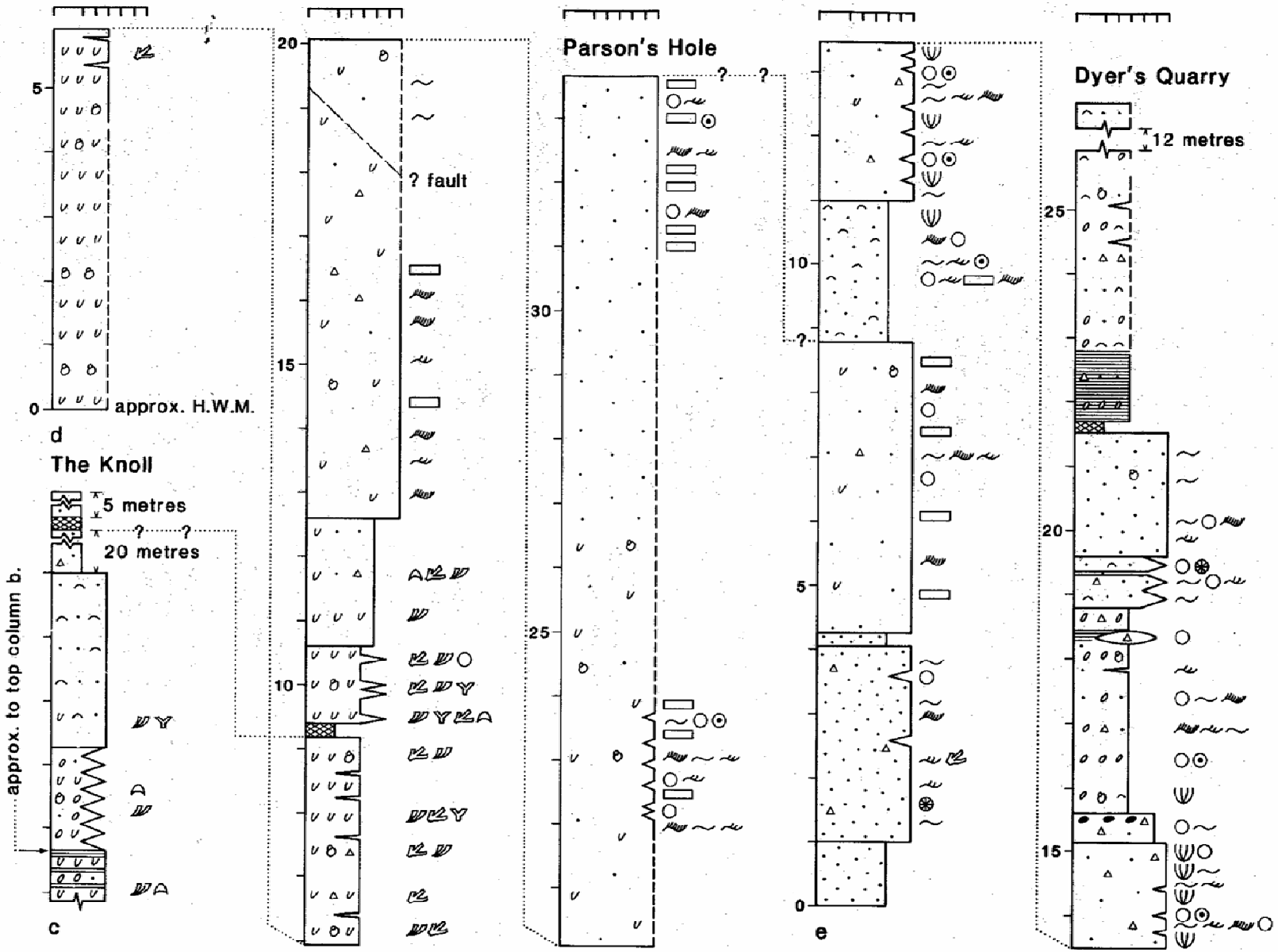

Figure 3. Stratigraphic logs of the Daddyhole Limestone and their proposed correlation:
a. Daddyhole Cove; b. Triangle Point; c. The Knoll; d. Parson's Hole; e. Dyer's Quarry.

Localities shown in Figure 2. Column lengths with a dashed margin represent inaccessible sections. Zigzag margins represent interbedded units too thin for resolution at this scale.

Coralline faunas and faunal associations.

Faunal associations are recurrent groupings of a particular suite of genera and species. They represent elements of one or more communities that appear to have been related, either as sequential or alternating colonisers of broadly discrete habitats in time and space.

Tabulate corals

Tabulate corals dominate the coralline faunas of the Daddyhole Limestone in the stratotype section. 27 species assigned to 13 genera are recognised and diversity is comparable to that in similar Middle Devonian sequences across Europe. The taxonomy of these corals is currently under review by K.B.G. but many species identified here can generally be recognised with reference to Lecompte (1939) and Dubatolov (1959).

Volumetrically, auloporoids are possibly of greatest importance as identifiable contributors to bioclasts, as well as occurring as fasciculate and reptant colonies on bedding planes and as encrusters of other corals. Species of *Remesia* (Fig. 4d, g) are particularly common throughout the upper 20 m of the Triangle Point sequence. They may occur alone, but are frequently associated with *Scoliopora* and, less commonly, *Thamnopora*. However, the latter is a prominent constituent with *Remesia* and *Scoliopora* near the base of the Parson's Hole sequence. *,4ulopora* itself is small, scattered and may be easily overlooked. It is largely present as an encruster of other coralline fauna, principally the rugose Corals. *Mastopora compacta* (Chernyshev) is slightly less common and also mainly present as an encruster. *Syringopora* is only present as a symbiont of stromatoporoids (see below).

Thamnopora, although present throughout the limestone, is only abundant in the rudstones forming many of the biostromes of Triangle Point (except those in the *Remesia-rich* peloidal packstones) and those near the base of the Parson's Hole sequence (Fig. 4e). Six species occur, including *T. reticulata* (de Blainville), T. *nicholsoni* (Frech), *T. parvula* (Ross), *T. yanetae* (Dubatolov) and two new species; up to four species may be present in the same biostrome. Two broad associations can be defined. One, usually with species of *Alveolites*, sometimes abundant *Cladopora* and occasionally *Striatopora tenuis* (Lecompte) characterises the limestones for the lower 28 m of the Triangle Point succession. The other, with *Scoliopora* and *Remesia* has been mentioned above. *Striatopora tenuis* is not common except at one horizon and only occurs with *Thamnopora*. *Scoliopora*, which is present as *S. denticulata denticulata* (Edwards and Haime), *S. d. longispina* (Lecompte) and a new species, is abundant but strongly restricted to the range of its association with *Remesia* and *Thamnopora*, although it may occur within that range in beds accompanied by *Remesia* only.

None of these branching tabulate corals is found in a bafflestone, in contrast to the fasciculate rugosan *Thamnophyllum*. The tabulate colonies are fragmented and the broken branches, which may reach 150 mm + long, rarely 300 mm long, are prostrate. If they trapped sediment during life, then it was largely winnowed out at the time of death of the community.

Alveolites is a prominent constituent of faunas in the limestone only near the top of the Parson's Hole section and the base of the Dyer's Quarry section. Four species are recognised, *A. minutus* (Lecompte) and three new species. They form a distinct association with *Planocoenites gradatus* (Lecompte) and *Cladopora vermicularis* (M'Coy) (Fig. 4h). *Thamnopora* is occasionally present. There are two versions of this association. *,Alveolites* is more prominent in bindstones with tabular stromatoporoids in the horizons mentioned above. This level is both preceded in the mid-section of the Parson's Hole sequence and succeeded in the seaward cliff in the upright limb of the syncline in Dyer's Quarry by an association with the same tabulate coral elements but dominated by repeated thin laminae of *Planocoenites* and with stromatoporoids rare or absent. At the base of the limestone in Daddyhole Cove, *Alveolites* is prominent in shale interbeds between thin crinoidal packstones and in the shales below.

Massive globular and domal tabulate corals are rare. *Heliolites porosus* (Goldfuss) is present as scattered colonies in Dyer's Quarry and Parson's Hole with the *Alveolites* or *Planocoenites* associations. *Favosites* has not been recorded from the Daddyhole Limestone.

Five principal associations are thus recognised, named for convenience as follows: 1. *Remesia (Remesia* almost exclusively); 2. *Scoliopora (Scoliopora, Remesia*, sometimes *Thamnopora* which may be abundant); 3. *Thamnopora (Thamnopora, Alveolites, Cladopora*, rarely *Striatopora); 4. Alveolites (Alveolites, Planocoenites, Cladopora*, some *Thamnopora*, often associated with stromatoporoids); 5. *Planocoenites (Planocoenites, Cladopora*, other elements rare).

Rugose corals

Rugose corals are represented by 20 species, all but one solitary, belonging to 10 genera; the bulk of them are cystimorphs and ptenophyllids. Most have been identified with reference to the works of Birenheide (1961, 1964, 1972), although the generic division of the Digonophyllidae proposed by McLean (1976, as the Digonophyllinae) has been followed. Diversity is rather less than in comparable Middle Devonian facies in Belgium and Germany. It increases through the Daddyhole Limestone to a maximum associated with the *Planocoenites* bindstones of Dyer's Quarry.

Two crude solitary coral associations can be distinguished. A Ptenophyllid association contains either or both of A*canthophyllum* and *Grypophyllum*, with *Lekanophyllum*, *Cystiphylloides* and *Digonophyllum* variably represented. This association, which is poorly diversified and often sparsely represented, occurs in Daddyhole Cove and up to 35 m above the base of the Triangle Point succession. In the 5 m transition zone at the base of the Daddyhole Limestone and the immediately overlying crinoidal grainstones it is joined by *Calceola sandalina*. *Calceola* appears to be restricted to this horizon in the Torquay Limestone Group. The faunas up to 17 m on Triangle Point are mainly associated with the thin biostromes of *Thamnopora*

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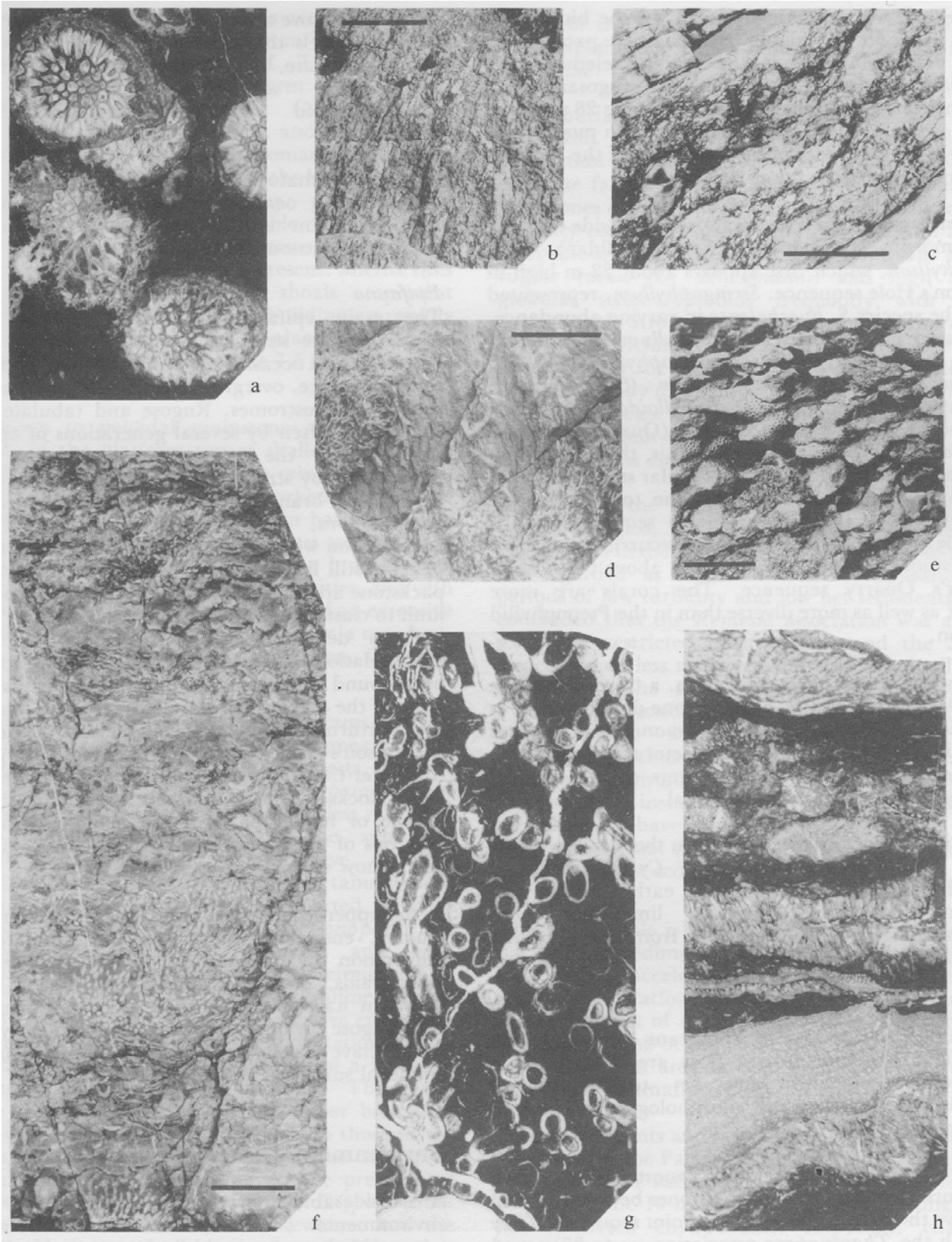

Figure 4. (a) Stromatoporoid encrustation of *Thamnopora sp.* nov. Triangle Point, 1 m thick biostome 28 m above base. x 2-5.
(b) Aligned *Mesophyllum laeve.* Quarry floor in Dyer's Quarry, 15 m above base. Scale 500 mm.
(c) 1 m thick biostrome. Triangle Point, 28 m above base. Scale 500 m.
(d) Colonies of *Remesia tubaeformis* (Goldfus). Parson's Hole, 10.5 m above base. Scale 100 mm.
(e) Biostrome of prostrate branch fragments of *Thamnopora reticulate* (de Blainville). Parson's Hole, 10 m above base. Scale 100 mm.
(f) *Thamnophyllum* bafflestone, *Planocoenites* bindstone and solitary coral rudstone/floatstone. Dyer's Quarry, 12 m above base. Scale 100 1m.
(g) *Remesia tubaeformis.* Parson's Hole, 11 m above base. x 2.5
(h) *Planocoenites gradatus* laminae with associated branch fragments of *Cladopora vermiculans* and a bryozoan. A tabular stromatoporoid is interleaved below and *Alveolites sp.* caps the sequence. Parson's Hole, 22-24 m above base. x 2.5.

In each case, base refers to the base of the sections logged in Figure 3

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rudestone and are numerically weak. In the biostromes scattered through the higher *Remesia-rich* packstones, however, *Grypophyllum* is the leading element. G. *denckmanni* Wedekind dominates the rugosans as a prominent component of the thick biostrome 28 m above the base of thee Triangle Point succession. In most of the higher beds, and in those in the lower part of the Parson's Hole succession, rugose corals are rare.

The second association is characterised by a wide diversity of cystimorphs (principally Digonophyllidae), with *Stringophyllum,* which first appears about 22 m high in the Parson's Hole sequence. *Stringophyllum,* represented only by the species *S. buechelense* in varying abundance, is associated chiefly with *Lekanophyllum cylindricum,* although *L. annulifer* (Schlüter), *Mesophyllum laeve, M. maximum* (Schlüter), M. *critatum* (Schlfiter), M. *lissingenense* (Schlüter), *Cystiphylloides limbatum* (Quenstedt), *C. antilimbatum* (Quenstedt) and *Ancanthophyllum* ?nay also occur. This, the Cystimorph association, is a component of the tabular stromatoporoid and *Alveolites* bindstones towards the top of Parson's Hole and low in the Dyer's Quarry sequence, but is principally developed as frequently recurring rudstones and floatstones between 10 and 20 m above the base of the Dyer's Quarry sequence. The corals are more abundant as well as more diverse than in the Ptenophyllid association.

In the lower 6 m of this same section, a third distinctive rugosan association occurs, a bafflestone dominated by *in situ* thickets of *Thamnophyllum germanicum schouppei* Scrutton in colonies up to 1.5 m diameter and 0.4 m high. Isolated examples and smaller clusters of this colonial rugose coral also occur in this interval.

The distribution of the cystimorphs in the Tor Bay Reef-Complex is interesting. Although *Cystiphylloides* is present throughout the Eifelian and early Givetian pre-reef and platform interior limestones, the Digonophyllidae are recorded only from the marginal areas of the carbonate platform.

Stromatoporoids

The stromatoporoids are an important element in the Daddyhole Limestone faunas. They are of variable to poor preservation and have yet to be studied taxonomically, but they fall morphologically into two broad groups.

Domal .to vertically discoidal coenostea are major components of rudstones and floatstones between 12 and 35 m above the base of the Triangle Point sequence. They occur with the *Thamnopora* association up to 28 m and the *Scoliopora* association above. Throughout, the rugose corals belong to the Ptenophyllid association. There is a strong correlation between the *Scoliopora* association and the presence of symbiotic *Syringopora* (the caunopore state) in a high percentage of the coenostea. In almost all cases the stromatoporoids are disoriented.

Tabular stromatoporoid coenostea are scattered components of biostromes in the lower 12 m of the Triangle Point sequence. They are best developed, however, in the upper levels of Parson's Hole and between 4 and 9 m above the base of the Dyer's Quarry sequence. At these levels the stromatoporoids, some showing the caunopore state, are associated with the *Alveolites* and Cystimorph associations forming *Alveolites*-stromatoporoid and stromatoporoid dominated bindstones.

Digitate stromatoporoids are less common. *Stachyodes* occurs as an occasional component of biostromes on Triangle Point and is only prominent in the thick rudstone-floatstone at 30 m.

Epifauna

The main epifaunal encrusters in the Daddyhole Limestone are stromatoporoids, auloporids, *.4lveolites, Chaetetes* and occasionally *Planocoenites.* In the Triangle Point sequence, overgrowth is usually common except in the lowest biostromes. Rugose and tabulate corals are overgrown, often by several generations of epibionts. In many cases, the even nature of the overgrowth particularly by stromatoporoids, completely around, for example, branches and branch junctions of *Thamnopora,* or *Grypophyllum* coralla, suggests encrustation whilst the coral was still in growth position, possibly still living (Fig. 4a). Bioclastic fragments in the packstone are uncoated, but there may have been a lower limit to clast size for successful colonisation by encrusters, perhaps depending on the rate of fine sediment accumulation. Rare rugose coralla at other horizons have been found encrusted post partial decortication. Thus some of the overgrowth was definitely post-mortem, after the overturning and at least slight wear of the skeletons. In Parson's Hole, successive laminar overgrowths of *,41veolites-Chaetetes-Stromatoporoid* have been found in loose blocks which presumably come from inaccessible middle or higher parts of the sequence. In accessible outcrops of the bindstones at these levels, rugose corals again show extensive overgrowth.

In the upper levels of the Daddyhole Limestone, in Dyer's Quarry, encrusting of rugose and tabulate corals is uncommon by comparison. Occasional examples of auloporaids, *Chaetetes* and in one case a fistuliporid bryozoan have been recorded as overgrowths. In some cases rugose coralla are decorticated and early encrusters could have been destroyed, but often uncolonised epithecal surfaces can be seen.

Environmental interpretation

A considerable literature is available on carbonate environments both ancient and modern. Recent summaries drawn on in this interpretation include Wilson (1975), Bathurst (1975) and Shinn, Enos and Wilson and Jordan, all in Scholle *et al.,* (1983). Single criteria, with the exception of desiccation cracks, tend not to be strongly diagnostic of particular environments of deposition and we have relied heavily on the juxtaposition of faunal and petrographic evidence in vertical sequence.

The base of the Daddyhole Limestone is marked by the first tongue of crinoidal debris sheet-transported into an area of background mud deposition from the migrating margin of a carbonate platform. Throughout the Tor Bay

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Reef-Complex, the basal horizons are crinoidal and vary in age from place to .place (Scrutton 1977a, b). The crinoidal bioclastic grainstones of Triangle Point represent the arrival of the platform edge proper, dominated by groves of crinoids. Colonisation by coralline faunas, the *Thamnopora* and Ptenophyllid associations with non-caunopore stromatoporoids, was episodic and indeed, some material in the biostromes may represent transport from near-by platform top environments. There is little identifiable coralline debris in the grainstones, although bryozoan, brachiopod, gastropod and ostracod bioclasts are present. Micrite rims are common. These crinoidal shoals represent accumulation at or above wave base, building up rapidly and temporarily to intertidal level in places as indicated by the desiccation cracked surface at 7.5 m on Triangle Point.

The transition to peloidal grainstones represents the stabilisation of the platform top at shallow subtidal or intertidal level. Graded beds of fine crinoidal material were swept in and probably reworked across the depositional surface. Rare channel-form bases occur on some beds. The indigenous fauna appears to have been gastropods, bivalves, ostracods and some brachiopods, often showing geopetal fill or umbrellar cavities, and with peloids possibly representing faecal pellets of an extensive soft-bodied biota. Possible charophyte oogonia may have been derived from a nearby brackish lagoon. Coralline material is very rare.

Above, intraclasts and occasional low angle cross-bedding indicate shallow channel activity. The appearance of *Remesia* as scattered partial colonies and clasts, together with bioclastic material including *Thamnopora*, suggest either a reduction in sedimentation rate or less restricted conditions allowing colonisation of the substrate. Material in the two thick biostromes above, although dominated by disoriented stromatoporoids, tabulate and rugose corals, may not have been transported far (Fig. 4c). Ramose tabulates, principally *Thamnopora*, are present as large, branched fragments and local colonisation in conditions of more open circulation is indicated by the diversity of the coralline fauna, particularly the tabulate corals. The extensive encrusting of the corals suggests low rates of sedimentation which, with continuing subsidence, may have been a factor in establishing more favourable conditions. The lower biostrome and the lower part of the upper biostrome contain faunas essentially similar in kind to those in the biostromes of the grainstones lower in the sequence. Despite this overall diversity, however, the presence of one dominant species of rugose coral, *Grypophyllum denckmanni* which is not common elsewhere, and the apparently wide range of tolerance of most of the other principal elements, suggests that this may still have been a slightly abnormal environment. In contrast, the upper part of the higher biostrome contains the *Scoliopora* association, abundant *Remesia* in part, less common rugose corals and a high percentage of the domal and discoidal stromatoporoids showing the caunopore condition. This suggests a subtle change of environment and marginally more stressful conditions to judge by the distribution of these faunal elements at higher levels. In each case, masses of sediment introduced by storm surges may have caused the disruption, local transport and burial of the communities, perhaps episodically, which may explain the crude faunal zonation of the upper biostrome.

Final burial led to the re-establishment of intertidal conditions in which peloidal packstones contain desiccation cracks at two levels and a very restricted, non-coralline fauna. The succeeding bioclastic and peloidal packstones contain thin biostromes with a distinctive, low diversity coralline fauna of the *Remesia* and *Scoliopora* associations with very scattered Ptenophyllid association rugose corals *(G. denckmanni)* and high domed to globular stromatoporoids frequently with symbiotic *Syringopora*. Stromatoporoids may overgrow each other, often sandwiching an earlier generation of other encrusters on the surface of the dead coenosteum. The fauna is usually disoriented but with little fragmentation of more delicate material. Colonisation of a normally protected, non-emergent but somewhat restricted environment is envisaged, possibly a lagoon.

In general, the presence of stromatoporoids lacking ragged margins, high levels of encrusting and lack of bafflestones in the lower part of the Daddyhole Limestone are regarded as characteristic of colonisation during periods of low turbidity and low net sedimentation. It is concluded that the *Remesia* association was adapted to somewhat restricted environments and the *Scoliopora* association to less restricted but still marginally stressful conditions. The more widespread *Thamnopora* and Ptenophyllid associations seem to range from normal open marine situations into slightly restricted environments, thus showing a wider tolerance range. The balance of environmental parameters imposing stress in any particular instance is not clear but important factors are likely to have been low oxygen partial pressures and/or abnormal salinity or high salinity ranges, as well as conditions of sedimentation.

The Knoll Shale is interpreted as an intrusive tongue of background sedimentation introduced as the result of local, slightly accelerated subsidence of the margin of the carbonate platform, perhaps as the result of subcompaction of the Eifelian shales. The sequence on the upright limb of the Daddyhole anticline is considered to represent an area farther from the platform margin and therefore unaffected by this episode.

Similar sediments and faunal associations characterise the lower part of the Parson's Hole succession as the top of the Triangle Point succession. *Thamnopora* becomes more common some 10 m above the base suggesting a slightly less restricted environment. Important changes are indicated, however, by the incoming of entirely new faunal associations midway through the Parson's Hole sequence. The *Planocoenites* bindstones and solitary coral floatstones and rudstones containing the Cystimorph association, with increased coralline species diversity and absence of the characteristically high stress species seen lower in the limestone, are taken to indicate a return to normal open marine circulation. This faunal association is explored further in the discussion of the Dyer's Quarry sequence. The increasing size and incidence of tabular to low domal stromatoporoids above, together with the *Alveolites* association and the high

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percentage organic component of the rock, may represent the build-up of a bioherm on the platform top, on the *Planocoenites* foundation. The top of this facies is dominated by large tabular stromatoporoids, some of which show the caunopore state. Relief cannot be demonstrated but the faunal sequence is similar in general to that described by Lecompte (1968, 1970) and Tsien (1971) from Belgian bioherms and noted in many other build-ups of Devonian age elsewhere (Klovan, 1964; Embry and Klovan, 1971, for example). There is little indication of really high energy conditions, however, although some coenostea are disoriented. The characteristics of Lecompte's "zone turbulente" are missing and the internal packstone to wackestone matrix lacks coarse bioclasts. The sequence may represent autogenic succession rather than an increasingly turbulent environment (Walker and Alberstadt, 1975).

The similar but thinner beds in Dyer's Quarry have been mentioned as the possible flank of the Parson's Hole stromatoporoid bindstone facies, possibly here in slightly deeper water. It is succeeded by the highly distinctive, crudely cyclic sequence of *Thamnophyllum* baffiestones (poorly developed in the lower cycles), with alternations-of *Planocoenites* bindstones and solitary coral rudstones-floatstones each normally in the range 50-150 mm (Fig. 4F). The cycles are about 300-800 mm thick. Stromatoporoids completely disappear from the sequence at this level. There are similarities to the mid section of the Parson's Hole sequence but for the absence of stromatoporoids, the presence of *Thamnophyllum* and the higher diversity of the Cystimorph association. It represents a return to similar, probably more turbid, conditions with rapid sediment accumulation still in fully marine conditions and generally below wave base, as reworking episodes are limited to the formation of the rudstones and floatstones. *Thamnophyllum* is present as more or less uncompacted colonies in position of growth, which probably both trapped sediment in life and were eventually killed by blankets of sediment not subsequently reworked. *Planocoenites* repeatedly colonised sediment surfaces during brief periods of minimal sediment input, forming undulating laminae regularly 2-4 mm thick, smothered by a fine bioclastic wackestone to packstone sediment layer 2-15 mm thick. Estimated growth rates of Palaeozoic tabulate corals suggest that these may be annual growth laminae buried by sediment mobilised by winter storms (Scrutton and Powell, 1979). Very little sign of reworking of *Planocoenites* is seen, although the associated, delicate *Cladopora* and bryozoa are partly fragmented and prone in the sediment blanket. Sequences of the bindstone must represent almost undisturbed accumulations of sediment and fauna at a mean rate of approximately 10 mm yr-1. The solitary corals of the rudstones to floatstones are now almost all recumbent. Some are geniculate, showing continuing growth after disorientation. A very few corallites clearly grew prone, stabilised on the substrate by talons and with a calice oblique to the axis of growth and angled upwards; their cross-sectional shape tends to be irregular. The lack of such features in the bulk of the long, straight, cylindrical coralla suggests that they originally grew more or less vertically supported by soft sediment around their bases. This must have been of substantial thickness to support slender cylinders up to at least 600 mm long. It is

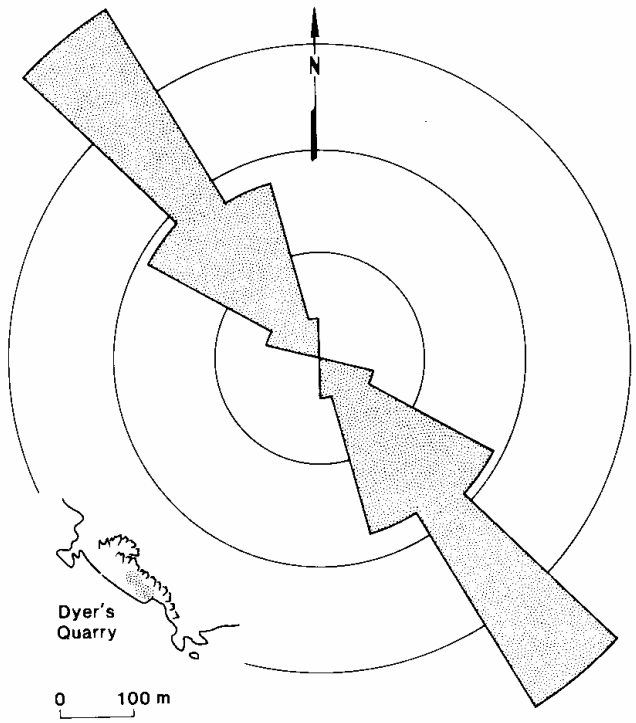

Figure 5. Alignment of solitary rugose corals on the floor of Dyer's Quarry. Measurements of 80 straight coralla, plotted in 15° classes. Circles indicate 10, 20, 30 observations. Area from which data collected shown in inset.

envisaged that occasional storms winnowed this supporting material, to topple them, cause the death of the majority and produce the remarkable alignment of coralla seen in the three bedding planes on the quarry floor in Dyer's Quarry (Figs. 4b, 5). Some alignment may also be present at lower levels but is not obvious on the available vertical faces. The relative rarity of epifauna on the epithecae of these corals suggests rapid covering by sediment to limit the availability of this substrate for colonisation.

The Cystimorph association is judged to be the most tolerant of high rates of sedimentation and probably accompanying turbidity, followed by *Thamnophyllum.* Also *Planocoenites* may be more tolerant of turbidity than *Alveolites* as their distribution, and the generally thicker colonies of the latter, would indicate. At least some cystimorph elements, however, were very broadly adapted in this sense, as they are equally successful in the *Alveolites* and stromatoporoid bindstones.

The microbioclastic packstones in the back wall of Dyer's Quarry, with some reworked sediment, peloids and a decreasing coralline fauna, represent a return to a somewhat restricted, perhaps slightly shallower environment. There was a brief return to conditions favouring the *Planocoenites* bindstones and solitary coral floatstones but following these, from the level of the tuff, beds which are sometimes laminated, with bivalves and gastropods, totally lack a coralline fauna. The absence of the *Remesia* association from these beds may reflect a higher sedimentation rate successfully inhibiting spat settlement or smothering early growth. The lack of clearly stenohaline forms as larger bioclasts, with the presence of fine crinoidal, peloidal and bioclastic material and a

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significant sedimentation rate suggests a restricted, lagoonal environment..

Different coralline associations were dearly characteristic of different environments, showing varying degrees of tolerance to a range of variables, probably including salinity, oxygenation, wave energy, turbidity and rate of sediment accumulation. A precise equation between species associations and specific environmental variables is speculative. Variation in faunal composition in detail, furthermore, will probably reflect patterns of colonisation of the substrate depending on many factors such as spat supply and settling success, interspecific competition and short term environmental stability. The utility of the associations we have recognised will depend on their applicability to other Old World Realm Middle Devonian carbonate shelf developments.

Broadly similar work by Brett *et al.* (1983) on Middle Devonian coral associations of New York State examined a somewhat different range of open shelf environments. None of their associations is rich in stromatoporoids and none is associated with clear evidence of intertidal conditions. Their associations are more broadly drawn and occur in more mud rich facies than those of the Daddyhole Limestone Formation. Comparisons are also less obvious because of the high endemicity of the Eastern Americas Realm in the Middle Devonian (Oliver 1977). However, their large Rugose Coral Association has some similarities with the Cystimorph plus *Planocoenites* associations herein although contrasts in detail in the characters of the faunas and their preservation reflect the differences inferred in their environments of growth. It should also be noticed that quite distinct auloporid associations characterise different types of more marginal environments in the two cases. Their *.4ulocystis* Association is inferred to be characteristic of deeper shelf argillites whereas the *Remesia* association herein is regarded as occurring in very shallow, possibly in part lagoonal environments.

Conclusions

The stratotype section of the Daddyhole Limestone represents the migration into the area of a carbonate platform largely constructed of crinoid debris. The platform rapidly built up to shallow water (intertidal to immediately subtidal) and was colonised by a range of coralline faunas from time to time. Intertidal and lagoonal facies in the lower part of the succession, generally with high stress faunal associations, are capped on the margin of the platform by a wedge of shales. Elsewhere, relative subsidence of the platform allowed fully marine faunas to colonise and to construct a substantial stromatoporoid bindstone, a possible bioherm, in conditions of low sediment input. Increased and in part episodic sediment influx in a subtidal environment excluded stromatoporoids but preserved a diverse fauna of coral pavements and thickets. Towards the top of the limestone a moderately high rate of sedimentation continued in a more restricted environment, perhaps a lagoon, effectively excluding a coralline fauna.

Several coralline associations are defined. Among the tabulate corals, the *Scoliopora* and *Remesia* associations are considered to have been adapted to successively higher stress environments. The *Thamnopora* association had a wide tolerance. The *Planocoenites* and *Alveolites* associations were generally fully marine, with the latter probably less tolerant of turbidity. The Ptenophyllid association among the rugose corals had the higher tolerance of restricted environments, although not as extreme as those colonised by *Remesia.* The Cystimorph association occurred in a range of fully marine facies and may have been tolerant of high turbidity but not of restricted conditions. Small domal, globular or vertically discoidal stromatoporoid coenostea, frequently in symbiotic intergrowth with *Syringopora,* were characteristic of more restricted environments. Tabular and domal coenostea also occur with faunas interpreted as fully marine in conditions of low sedimentation; these may also show intergrowth with *Syringopora.* They reach their acme in size and dominance in what is thought to have been the least turbid environment. Conversely, they are completely absent from environments considered to have had high levels of turbidity and rates of sediment accumulation.

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Read at the Annual Conference of the Ussher Society, January 1984

Ceriopora ramulosa (MICHELIN); an aberrant bryozoan from the Cenomanian of S. E. Devonshire

M. B. HART
K. JOHNSON

Hart, M. B. and Johnson, K. 1984 *Ceriopora ramulosa* (MICHELIN); an aberrant bryozoan from the Cenomanian of S. E. Devonshire. *Proceedings of the Ussher Society, 6,* 25-28.

Newly-collected specimens of *'Ceriopora ramulosa* (MICHELIN)' from the Cenomanian Limestone (Bed Al) of Beer Head have been investigated. These rather enigmatic fossils, described as the largest bryozoan in the British Isles, can be shown to be more appropriately regarded as sponges. They display many of the characteristics that indicate that they are either members of, or closely related to, the Class Sclerospongiae of Hartman and Goreau (1970).

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Introduction

One of the most distinctive features of Bed A1 of the Cenomanian Limestone succession in S. E. Devonshire is an abundance of fossils, especially where the bed is thickly developed. One such locality is Beer Head; a succession described in detail by Smith (1957a), Kennedy (1970), Carter and Hart (1977), and Hart (1982). At Beer Head, especially just to the west of the point (near the old sewerage outfall Pipe), the most distinctive feature of Bed A1 is the presence of the 'large coral-like branching polyzoan, *Ceriopora ramulosa* (MICHELIN), first described in detail by Smith (1957a). Smith (op.cit.) described these fossils as large colonial structures, rarely seen in position of growth, and generally bored by *Lithophaga* and smaller *'clionid'* sponges.

Smith's identification and interpretations of these fossils seem to have come from Dr. H. Dighton Thomas, who was present on a Geologists' Association Field Excursion in August, 1955; subsequently reported by Smith (1957b). Dr. H. Dighton Thomas identified the material as *Ceriopora ramulosa* (MICHELIN), and claimed that it was at that time the "largest polyzoan found in British rocks". This fossil was originally described from the Cenomanian of France under the name *Chaetites ramulosus* MICHELIN, but was later assigned to the bryozoan genus *Ceriopora* GOLDFUSS (see Gregory, 1909, p,167). Unfortunately the material collected from Beer Head is not like any of the other species of *Ceriopora* described from the U.K. and housed in the collections of the British Museum (Natural History) -- see Gregory (1899/1909). The senior author has looked at the collections in the museum, with the help of Dr. P. D. Taylor, and it is clear that none of the other species of *Ceriopora* are related in any way to the material under discussion. These collections were not discussed by Smith in his 1957a paper, or in any of his Other publications on the Devonshire successions (see Carter and Hart (1977) for full reference listing).

During the last two winters (1981/2 and 1882/3) several rock falls at Beer Head have made accessible a large amount of new material, much of which is in a reasonable state of preservation. Many of the specimens show good surface features, while others are strong enough to allow the cutting of thin sections. Some of the material is shown in Fig. 1 a-g.

Description of the Material

The material is currently housed in the research collections of Plymouth Polytechnic, although it is hoped that a suite of specimens may eventually be deposited in the British Museum (Natural History). The material consists of a large number of fragments, but also several large tubular forms (5-10 cm long and 2-5 cm in diameter) together with a few, large, branching forms (6-10. cm .long). Two palmate growth forms have been seen m the field, but their collection was not attempted. The majority of specimens are broken, abraded, and definitely *not* in position of growth. No basal structure, holdfast, or point of attachment has been seen, either in the collected material, or on some of the larger specimens left in the field. All the material is affected by boring organisms, and many of the structural details of the fossils are concealed by these infilled tubes and holes. These structures can be summarised as:-

1. Large, *Lithophaga-like,* bivalve borings that are infilled with coarse grained, glauconitic, sandy, chalk (Fig.la,lc). These must certainly be post-mortem, as they go directly through the whole specimen, although some recent corals and bryozoans can tolerate extensive boring (especially in basal regions).
2. Medium-sized *Cliona-like,* boring/tubes, ofen infilled with rather variable, fine-grained sediments (Fig. le, -lg).
3. Very fine, rather irregular, tubes (often iron stained),

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Figure 1. Sclerosponges of Early Cenomanian age from the Cenomanian Limestone (Bed A1) of Beer Head, S.E. Devonshire.

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that are normally restricted to the margins of the specimens (see especially the right hand side of Fig. lg).

The outer surface (only shown in Fig. 1f) appears to be covered in a complex set of polygonal pits (calicles?), although this is rarely seen as the majority of specimens are badly preserved and/or covered in rather pervasive sediment. Although difficult to identify on Fig. 1f there are three or four features that give the impression of weakly developed astrorhizae, a morphology not seen in the Phylum Bryozoa. Very similar features are recorded by Hartman and Goreau (1070, fig. 14) in their description of modern coralline sponge material from Jamaica. If these surface pits are true calicles, their scale is entirely comparable to many modern sclerosponges (see Hartman and Goreau 1975, pl.5). Our specimen (Fig. If) also shows a few small pits between some of the calicles, and these could represent asexually produced buds, also described by Hartman and Goreau (1975, pl.5). Without detailed SEM photography it has not been possible to identify any structures within the calicles, but there is some evidence to suggest that internal spines are present in some cases (as shown by Hartman and Goreau, 1975, pl. 5,6,7). These are also features not seen in any bryozoans.

Below the surface layer the calicles appear to taper rapidly (Fig. 1a-c, d, g) and become infilled with CaCO3 (Fig. lb, g). The whole centre of the specimen thereby becomes a solid mass of CaCO3 and further detailed SEM work will be needed to interpret the visible structures. No siliceous (or carbonate) spicules have been seen in thin section, or in the acid reductions of some fragments attempted during the present phase of the research.

Affinities

If, as it seems likely from the above, these strange fossils are not bryozoans, then the sponges are clearly the most likely taxonomic group for comparable forms. The presence of astrorhizae and the general external and internal appearance points to the Class Sclerospongiae of Hartman and Goreau (1970). The sclerosponges from Jamaica described by Hartmand and Goreau (op. cit.) all belong in the Family Ceratoporellidae--which has the following characteristics:-

> "……sponges dwelling on coral reefs in habitats of reduced light such as caves, overhangs, and tunnels. The living tissue is organised as in typical Desmospongiae."
>
> "Skeletal elements include siliceous spicules, proteinaceous (probably collagenous) fibres and a basal mass of aragonite .... "
>
> "In all forms the cavities on the surface of the aragonitic basal mass are filled in by additional deposits of $CaCO_3$ as the skeleton grows upward so that the cavities retain a similar depth throughout the life of the organism."
>
> "The process of infilling by CaCO3 leads to the formation of a solid calcareous skeleton below the level of the surface processes."

Sokolov (1955, 1962)--as reported by Hartman and Goreau (1972)--has suggested that the chaetetids are probably closely related to the stromatoporoids. Hartman and Goreau (1972) go on to indicate that *Ceratoporella nicholsoni* (Hickson) would also appear to be very close to a chaetetid, but with one very big difference; chaetetids do not possess siliceous spicules. However Newell *et al* (1953) have suggested that in many ancient carbonates, siliceous spicules set in a carbonate matric tend to be replaced by $CaCO_3$. This could explain the lack of visible spicules in thin section and acid reductions. Detailed SEM work may, it is hoped, reveal 'ghosts' in the carbonate that could confirm this suggestion. Hartman and Goreau (1972) regard the chaetetids (range Ordovician to Cretaceous) as ancestral to the sclerosponges *sensu stricto.*

Palaeoecology

In their definition of the Ceratoporellidae, Hartman and Goreau (1972) suggest that the living tissue is organised in the same way as some typical demosponges. Recently Trammer (1979) has described the palaeoecology of a Jurassic demosponge, *Reiswigia ramosa* Trammer. This sponge reportedly changes its morphology in response to increasing water turbulence, and Trammer (op. cit.) has indicated that solid, flat, encrusting forms inhabited turbulent waters, while solid cylindrical (or solid branching) forms--as we have in our material--lived in slightly less turbulent waters. Trammer argues that they grew into an upright position, 8-10cm above the substrate, to maintain a water flow through their tissues.

Hartman and Goreau (1970), in a lengthy account of coralline sponges from Jamaica, have presented some interesting data on the distribution of living sclerosponges. On the Jamaican coastline (18°N of the equator) the sponges are most numerous on the open sea floor at depths of 70-95 m. *Ceratoporella nicholsoni* (Hickson) does inhabit shallower waters but only when shaded. At depths of 30-40 m it lives exclusively in caves and tunnels. Many of the other species from Jamaica are highly photophobic, living only in cave systems along the coastline. It is interesting that one species, *Stromatospongia vermicola* Hartman, is described as having an 'intimate relationship with an unidentified serpulid worm'. It should be possible, with further work, to decide if the finer borings and tubes seen in the Devon material are pre-death or post-death.

If, as indicated by Smith (1957a), these sclerosponges were living attached to the substrate, and were then swept by increasing turbulence into troughs that now represent the thicker developments of Bed Al, what can one say about the water depths involved? Depths of 70-95 m would appear to be excessive, both for the turbulence, and the close stratigraphical proximity to features such as the 'cobble conglomerate' which is clearly a shallow water feature (even if not quite so shallow-water as the beach

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deposit suggested by Ali (1976)). It is however difficult to imagine the Lower Cenomanian cave and tunnel systems needed to provide the shade the sclerosponges now seem to require for them to inhabit shallower water depths. It is possible that this modern distribution is the result of competition from other organisms (scleractinian corals?) and that their Cretaceous depth distribution would have been somewhat different.

Conclusions

This preliminary investigation into the affinities of *Ceriopora ramulosa* (MICHELIN) from the Cenomanian Limestone succession has shown that these fossils are probably not bryozoans, and that the above name is inappropriate. They are provisionally regarded as sclerosponges (perhaps chaetetids), and until further work is undertaken they are not given any formal name. Eventually it may .be possible to use them, and their environmental limitations, to provide further evidence of the palaeoecology of the Cenomanian Limestone succession in S.E. Devonshire.

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Read at the Annual Conference of the Ussher Society, January 1984

Late Triassic palynomorph records from Somerset

G. WARRINGTON

Warrington, G. 1984. Late Triassic palynomorph records from Somerset. *Proceedings of the Ussher Society, 6,* 29-34.

Assemblages of miospores of middle (?) to late Triassic (late Ladinian (?) to Carnian) age have been recovered from beds around the level of the Somerset Halite Formation in the Mercia Mudstone Group succession proved in the Puriton Borehole near Bridgwater, Somerset. Palynomorph assemblages of latest Triassic (Rhaetian) age occur in Penarth Group deposits at Chilcompton, Holwell and Vallis Vale in the eastern Mendips.

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Introduction

Late Triassic (Carnian) miospore assemblages have been reported from the Mercia Mudstone Group succession proved in the Puriton Borehole (Warrington 1980) and from the North Curry Sandstone Member of that Group at outcrop near Taunton (Warrington and Williams 1984). Younger Triassic palynomorphs, of Norian (?) to Rhaetian age, have been recorded from the higher Mercia Mudstone Group to basal (pre-Hettangian) Lias succession on the west Somerset coast (Warrington 1974, 1978a, 1981, in Whittaker and Green 1983; Fisher and Dunay 1981; Warrington and Whittaker 1984) and in the Burton Row Borehole, Brent Knoll (Warrington 1980, 1981, *in* Whittaker and Green 1983), and from the Penarth Group at Vallis Vale (Orbell 1973). The miospore assemblages from the Puriton Borehole and palynomorph assemblages from Penarth Group deposits at Chilcompton, Holwell and Vallis Vale, in the eastern Mendips, are documented in this account.

The Puriton Borehole

Until the completion of the British Geological Survey Burton Row Borehole on Brent Knoll in 1972, the Puriton Borehole, near Bridgwater (Ussher *in* Richardson *et al.* 1911; McMurtrie 1912), was the only one in the Somerset region that proved a substantial sequence regarded as Permo-Triassic in age. In this account, the lithostratigraphic nomenclature used by Ussher and McMurtrie is replaced by that introduced by Warrington *et al.* (1980); thicknesses and depths are from the account by McMurtrie *(op. cit.),* and from extant samples, and have been metricated.

The Puriton Borehole (site ST31913083) was drilled between December, 1909 and November, 1910. Beneath alluvium, the borehole entered the Triassic succession slightly below the top of the Mercia Mudstone Group; 381.7 m of beds referable to that group, and including a halite-bearing unit some 22.2 m thick (the Somerset Halite Formation) were proven overlying 65.5 m of sandstones and conglomerates of the Sherwood Sandstone Group. The borehole continued in argillaceous sandstones now assigned to the Aylesbeare Group and was terminated in those beds at a depth of 631.6 m. It was cored throughout and 598 m of core, a recovery of nearly 95%, was obtained.

In the absence of biostratigraphic information from the borehole, cores held in the collection of the British Geological Survey and at the Bristol City and the Somerset County museums have been examined for material suitable for palynological study. Twenty samples (Table 1), two from the Sherwood Sandstone Group and the remainder from the Mercia Mudstone Group (Fig. 1), were selected for this purpose, No suitable material was obtained from the Aylesbeare Group.

Table 1. Palynology samples from the Puriton Borehole.

Preparation number[1]	Source[2]	Depth (metres)
2452	B, T	15.24
2453	T	25.6
2454	T	30.18
2455	B, T	30.48
2456	B	53.34
2457	T	114.86
2458	B, T	121.87
2459	B	161.54
2460	B	176.02
2461	B	181.66
2462	B	186.54
2463	B	191.44
2464	B	199.39
2465	T	217.02
2466	T	235.31
2467	B, T	287.83
2468	B	508.76
2469	T	342.29
2470	B, T	437.21
2471	B	443.59

1. Preparations and slides are held in the micropalaeontological collections at the British Geological Survey, Keyworth, Nottingham, and are registered in the SAL series.
2. B -Bristol City Museum (accession number 45/1965)
   T - Somerset County Museum, Taunton.

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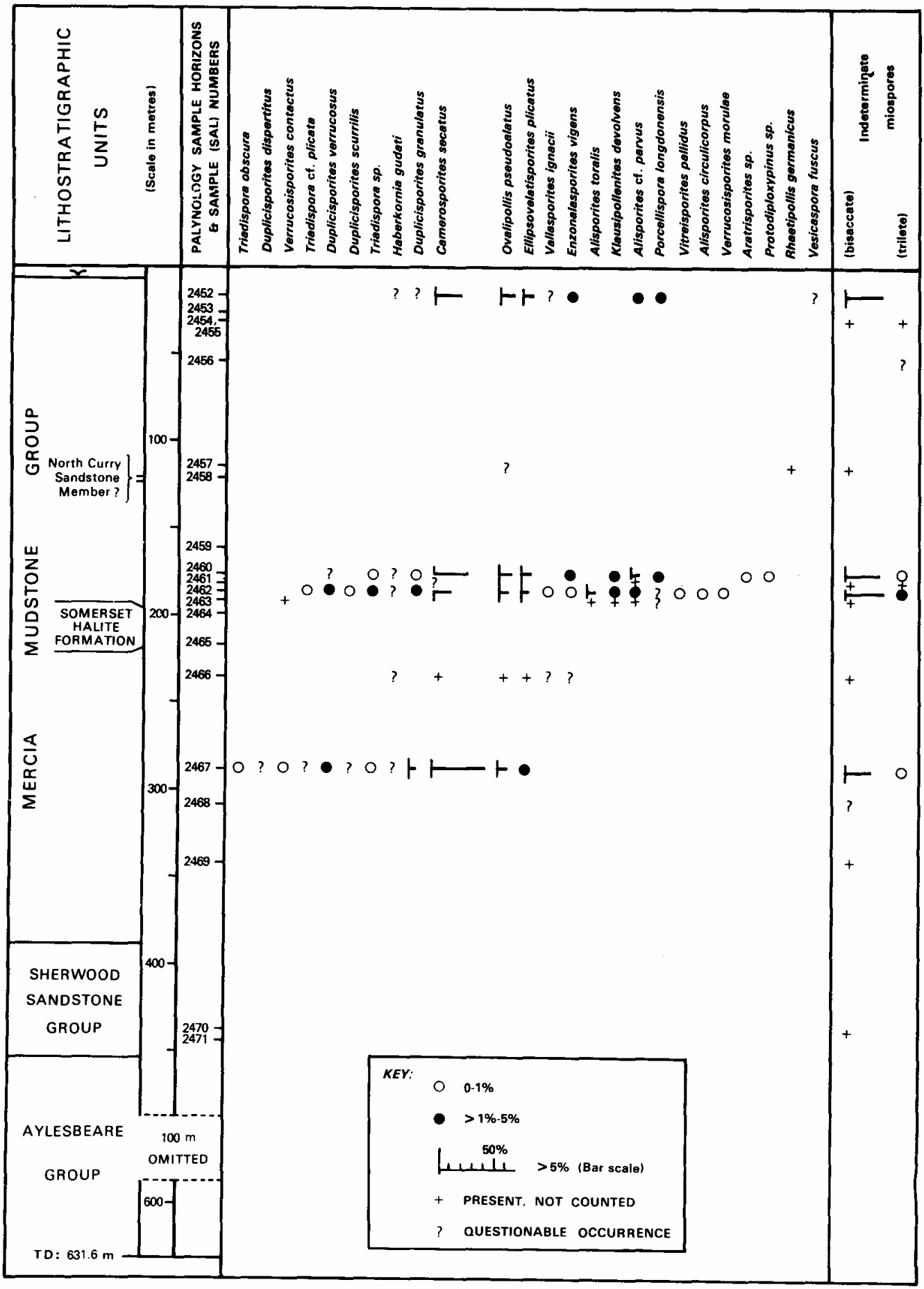

Figure 1. Stratigraphic distribution and composition of miospore assemblages from the Puriton Borehole, Somerset. Relative abundances are expressed as percentages based upon counts of 200 specimens.

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Palynological results

a. Sherwood Sandstone Group
Samples from this unit originated from the lower part of a sandstone sequence that overlies a thin conglomerate (McMurtrie 1912) and is correlated with the Otter Sandstone Formation. The preparations yielded fragments of dark brown and black organic material and a solitary indeterminate bisaccate miospore was recovered from 437.21 m (SAL 2470), This specimen, which is of post-Carboniferous age, is, in the context of the Puriton section, indicative only of a general Permo-Triassic age for these beds.

b. Mercia Mudstone Group
Of the 18 preparations from this unit, eight yielded determinable miospores and a further five yielded indeterminate remains; five samples, including those from the halite formation, were devoid of palynomorphs (Fig. 1).

Miospore assemblages from between 287.83 and 176.02 m, from beds below and immediately above the Somerset Halite Formation (SAL 2460, 2463, 2466, 2467; Fig. 1), are mostly sparse and comprise generally poorly preserved specimens. They are characterised by the presence of *Camerosporites secatus* Leschik *emend.* Scheuring 1978, *Duplicisporites spp., Ellipsovelatisporites plicatus* Klaus 1960, ? *Haberkornia gudati* Scheuring 1978, *Ovalipollis pseudoalatus* (Thiergart) chuurman 1976 and largely indeterminate disacciatriletes with, less consistently, *Triadispora spp; Verrucosisporites* occurs sporadically.

The principal assemblage from beds below the halite formation (SAL 2467) Comprises the above taxa and is dominated by *Camerosporites secatus* (Fig. 1). This is also an important constituent, with *E. plicatus, O. pseudoalatus* and disacciatriletes (including *Alisporites, Klausipollenites, Protodiploxypinus, Vitreisporites* and numerous indeterminate specimens) in assemblages SAL 2460 and 2462 from beds close above the halite formation, but is there associated with additional taxa, notably *Enzonalasporites vigens* Leschik 1955 *sensu* Scheuring 1970, *Porcellispora longdonensis* (Clarke) Scheuring 1970 *emend.* Morbey 1975 and *Vallasporites ignacii* Leschik 1955 *sensu* Scheuring 1970 (Fig. 1).

The miospore assemblages from between 176.02 and 287.83 m in the Puriton Borehole are, by comparison with those documented from other European successions for which independent biostratigraphic evidence is available, of middle (?) to late Triassic (Ladinian? to Carnian) age. Comparisons with palynological results from Ladinian and Carnian sequences in the Dolomites and Swiss Alps (Scheuring 1970, 1978; Mostler and Scheuring 1974; van der Eem 1983) and Sicily (Visscher and Krystyn 1978) indicate that the assemblages from 176.02 to 191.44m, which contain *Duplicisporites spp.* and *Vallasporites ignacii* in association with *Camerosporites secatus* and *Ovalipollis pseudoalatus,* are of Cordevolian or Julian (early or middle Carnian) age. The assemblages from beds below the halite formation lack definite specimens of *Vallasporites ignacii* and may be pre-Cordevolian but, because they lack *Echinitosporites iliacoides* Schulz and Krutzsch 1961, are unlikely to be older than Langobardian (late Ladinian).

The age thus assigned to beds between 176.02 and 287.83 m in the Mercia Mudstone Group succession at Puriton is slightly older than the Julian or Tuvalian (mid or late Carnian) date attributed to the North Curry Sandstone Member in the Group at outcrop to the east of Taunton (Warrington and Williams 1984). This is consonant with the presence of a possible representative of that member, comprising 3.1 m of "grey shale with markings", (McMurtrie 1912) between 121.6 and 124.7 m in the Puriton Borehole (Fig. 1); a sample (SAL 2458) from that unit proved devoid of palynomorphs.

Samples SAL 2452 and 2457 are from specimens reputedly from depths of 15.24 and 114.86 m respectively, but the miospores recovered (Fig. 1) are not consistent with that attribution. The assemblage from SAL 2452 is comparable with those recorded from between 176.02 and 191.44 m and is of Carnian age. In contrast, SAL 2457 yielded a very sparse association composed principally of *Rhaetipollis germanicus* Schulz 1967, a miospore indicative of a latest Triassic (late Norian? to Rhaetian) age. These samples may, therefore, have originated at lower and higher levels in the borehole respectively. Though probably displaced, the presence amongst the extant Puriton cores of material yielding R. *germanicus* is evidence for the existence of beds as young as Rhaetian in the upper part of that section and supports the view (McMurtrie 1912, p.49) that the borehole commenced as little as 23.0 m below the Penarth Group. Specimens of *R. germanicus* have been recovered from the highest 45 m of the Mercia Mudstone Group succession at St. Audrie's Bay, near Watchet, Somerset (Warrington 1974; Warrington and Whittaker 1984) and from the highest 52 m of that Group in the Burton Row Borehole, Brent Knoll, Somerset (Warrington *in* Whittaker and Green 1983).

Eastern Mendip localities

Palynological preparations (Fig. 2) have been made from Penarth Group deposits at the following localities:

1. *Chilcompton railway cutting* (ST653523)

This section, described by Richardson (1911, pp. 66-67) and Duffin (1980), is now obscured; samples were supplied by C. J. Duffin and represent the following horizons in the succession he recorded *(op. cit.,* p.255).

Lilstock Formation, Cotham Member:
SAL 6095:0.61 m below top of unit b, cycle Cb2
SAL 6094, base of unit b, cycle Cb2
SAL 6093:0.30 m below top of unit b, cycle Cb1

Westbury Formation:
SAL 6092:0.16 m below top of unit b, cycle Wb2

2. *Marston Road Quarry, Holwell* (ST730449)

This section, in a disused quarry, was recorded by Moore (1867, p.482) and Richardson (1911, p. 62) and was re-excavated and cleared in 1981/82 by the Nature

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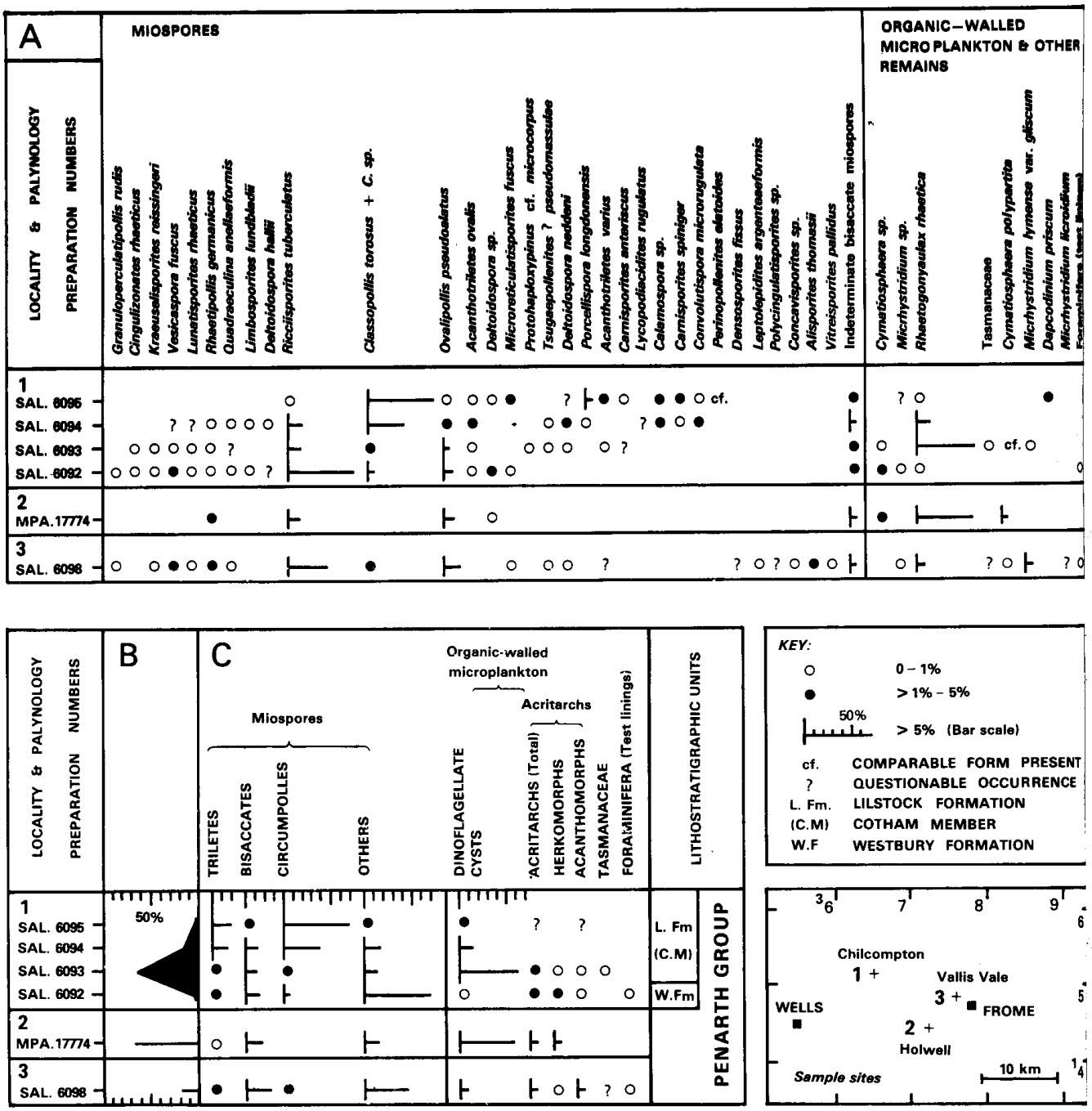

Figure 2. Palynomorph assemblages from Penarth Group deposits at Chilcompton (1), Holwell (2), and Vallis Vale (3). Relative abundances of miospore and organic-walled microplankton taxa and other remains (A), of organic-walled microplankton (black) to other palynomorphs (B), and of major groups of palynomorphs (C), are expressed as percentages based upon counts of 200 specimens. Preparations and slides are held in the micropalaentological collections at the British Geological Survey, Keyworth, Nottingham, and are registered in the SAL and MPA series.

Conservancy Council. A sample (MPA 17774), from black mudstones which rest upon Carboniferous Limestone and are attributed to the Penarth Group, was supplied by M. J. Harley (NCC, Geology and Physiography Section).

3. *Hapsford Bridge, Vallis Vale* (ST760494)

Orbell (1973 p. 27) briefly noted two palynomorph assemblages from a section in Penarth Group deposits about 90 m south-west of Hapsford Bridge. The succession here, recorded by Moore (1867, p.490) and Richardson (1911, p.65), has been revised by C. Copp (p.157 *in* Duffin 1982); samples examined by the writer are from the following beds and levels in the revised succession:

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SAL 6100: bed 7, 0.76 m above the base of the Penarth Group
SAL 6099: bed 6, 0.69 m above the base of the Penarth Group
SAL 6098: bed 5, 0.61 m above the base of the Penarth Group
Palynomorphs (Figure 2-3) were only recovered from SAL 6098

The palynomorph assemblages from the above localities comprise miospores and organic-walled microplankton; test linings of foraminifera occur sporadically in preparations from Hapsford Bridge and from the Westbury Formation at Chilcompton (Fig. 2: A, C). Miospores dominate several of the assemblages, but that from Holwell and the lowest from the Lilstock Formation at Chilcompton are dominated by organic-walled microplankton (Fig. 2: B), principally the dinoflagellate cyst *Rhaetogonyaulax rhaetica* (Sarjeant) Loeblich and Loeblich *emend.* Harland *et. al.*, 1975. The assemblages from Chilcompton and Vallis Vale are moderately varied in character, but that from Holwell comprises comparatively few taxa, particularly of miospores (Fig. 2: A).

Assemblages from the Chilcompton section are characterised by the presence of the miospores *Classopollis torosus* (Reissinger) Balme 1957, *Ovalipollis pseudoalatus*, *Ricciisporites tuberculatus* Lundblad 1954 and, in most instances, *Rhaetipollis germanicus*, in association with *Rhaetogonyaulax rhaetica*. They are comparable in composition with those documented from the Westbury Formation and Cotham Member succession elsewhere in southern and central Britain (Orbell 1973; Warrington 1974, 1977a, b, 1978b, 1982), and are of late Triassic, Rhaetian, age. The occurrence of *Ovalipollis pseudoalatus*, *Rhaetipollis germanicus* and *Ricciisporites tuberculatus* with *Rhaetogonyaulax rhaetica* in the assemblages from Holwell and Vallis Vale (Fig. 2: A) is indicative of a similar (Rhaetian) age and of a stratigraphic level comparable with those from Chilcompton.

The presence of organic-walled microplankton and the remains of foraminifera is indicative of deposition in an aqueous environment of marine origin. The Vallis Vale and Holwell localities are close to the depositional margin of the Penarth Group in the eastern Mendips, but the composition(Fig. 2: A) and character (Fig. 2: C) of the organic-walled microplankton associations from those sites compare well with those from Chilcompton and other more distal situations. Thus, proximity to the contemporary shore line does not appear to have significantly influenced the nature of these associations, and the physical characters of the aqueous environment they inhabited were probably relatively uniform over wide areas.

Acknowledgements. The writer is grateful to the authorities of the Bristol City Museum and the Somerset County Museum, Taunton; for making material from the Puriton Borehole cores available for palynological analysis, and for approval to publish the results. Dr. C. J. Duffin (Morden, London) and M.J. Harley (Nature Conservancy Council) kindly supplied samples from Chilcompton and Marston Road, Holwell, respectively and, with H. C. Ivimey-Cook and M. Mitchell (British Geological Survey), are thanked for their comments upon a draft version of this - account.

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Read at the Annual Conference of the Ussher Society, January 1984

The radioactive accessory mineral assemblage of the Carnmenellis Granite, Cornwall.

N.L. JEFFERIES

Jefferies, N. L., 1984 The radioactive accessory mineral assemblage of the Carnmenellis Granite, Cornwall *Proceedings of the Ussher Society 6,* 35-40.

In the Carnmenellis pluton two episodes of megacrystic biotite granite emplacement are recognised, followed by the later emplacement of fine-grained granites and microgranite sheets. During magmatic differentiation of the biotite granites and tourmaline microgranites Zr, Th and Ce act as 'compatible' elements and show a clear trend of decreasing values from the earlier to the later units. In contrast uranium remains at a constant high level throughout granite differentiation. Uraninite is the principal U bearing phase in all the granite types examined, although it is suggested that in the tourmaline microgranite sheets uranium is transported in an aqueous phase, rather than as part of a silicate liquid.
The radioactive accessory mineral assemblage (monazite, apatite, xenotime, zircon and uraninite) into which Zr, Th, Ce and U are strongly partitioned, has crystallised at an early magmatic stage within the megacrystic biotite granites.

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Introduction

The Variscan granites of south-west England occur as a series of plutons, which extend in a linear belt from Dartmoor to the Scilly Isles. The granite types within the major plutons have been described by Exley and Stone (1982) who identified earlier biotite granites (Types B and C) intruded by later lithium mica granites (Type E). Within the Carnmenellis pluton only the coarse - grained biotite granite (Type B) and the fine-grained biotite granite (Type C) are represented.

The granite plutons are geochemically highly evolved, being enriched in Li, F, B, C1, Rb, Cs, Sn, Pb and U relative to the average 'felsic igneous rock' of Vinogradov (1962). The biotite granites contain up to 35ppm U, with a mean content of 12.2ppm (Simpson *et al.,* 1976), more than four times the 'average' value given by Vinogradov (op cit).

Basham *et al.* (1982) investigated the distribution of uranium in the biotite granites and concluded that more than 60% is contained in low thorium uraninites. Oxidation of uranium and destruction of uraninite in the weathering zone leads to depletion of uranium in the granite. Weathered granite contains 2- 3ppm U which may be considered to be the level held in resistate accessory minerals such as monazite, zircon and apatite. The Carnmenellis granite was chosen for a detailed study of the radioactive accessory mineral assemblage owing to abundant quarry exposures from which samples below the zone of intense weathering could be collected, and because of the apparent lack of hydrothermal alteration over much of the pluton.

Originally mapped in detail by Ghosh (1934) the pluton has an annular form with a medium-grained megacrystic biotite granite (Type III) occupying its central part. This is considered to be intrusive into a coarser-grained megacrystic biotite granite (Ghosh Types I and II) which forms the outer part of the pluton (Figure 1). Granite types I and II were distinguished by Ghosh in the field on textural criteria, but more recent authors (Chayes, 1955; A1 Turki and Stone, 1978) have concluded that they are modally and chemically identical.

Two elliptical bosses of fine - grained granite appear to cut the megacrystic biotite granites (Fig. l). The megacrystic granites are also cut by thin sheets, up to 3 metres in width, of biotite or tourmaline microgranites, commonly referred to as aplites. The order of emplacement suggested for the Carnmenellis pluton is Types I and II granite, Type III granite, the fine-grained granite, and finally the microgranite sheets.

Experimental Procedure

To locate the position of radioactive accessory minerals in thin sections of granite, the technique of alpha autoradiography was employed. CR-39 is a commercially available nuclear track plastic (supplied by Pershore Mouldings Ltd., Pershore, Worcestershire) sensitive to ionizing radiations in the range 1-8 MeV; that is, the energy range of alpha-particles.

Sheets of CR-39 were positioned over uncovered thin sections and taped securely into place. After a suitable exposure time, which for this study of radioactive accessory minerals was approximately two weeks, the CR - 39 was removed from the thin section and etched to reveal the distribution of alpha- tracks. To etch the alpha - tracks the CR - 39 was immersed in 6.25N NaOH at a temperature of 75 °C for 6 hours.

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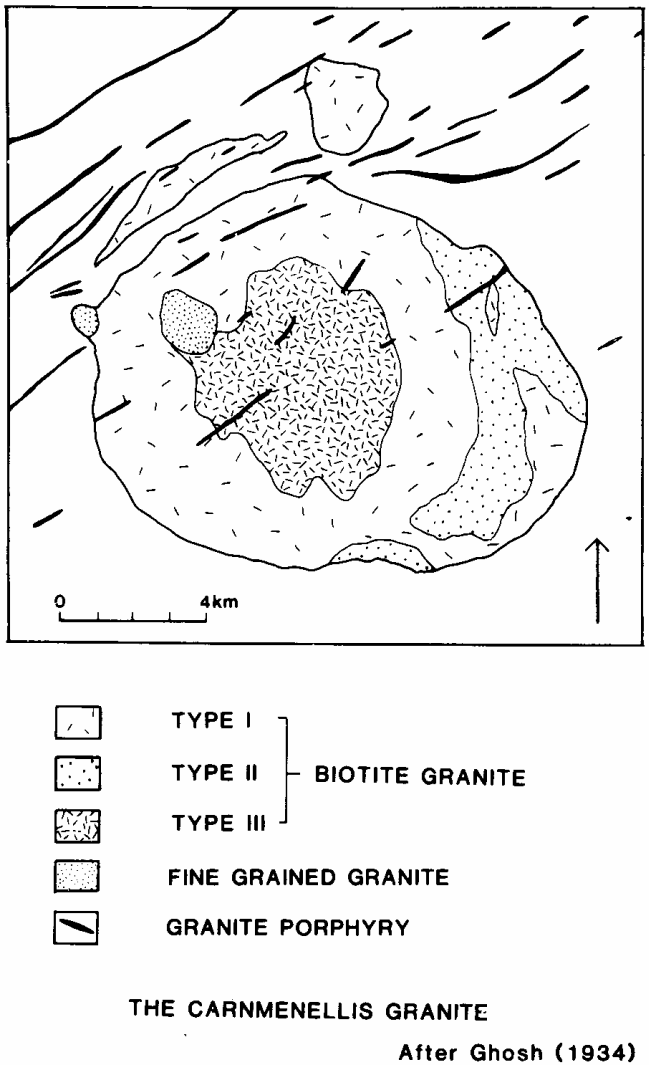

Figure 1. Outline geological map of the Carnmenellis granite

The radioactive minerals which were located by means of the alpha- autoradiographs were then examined using an energy dispersive Link microanalyser attached to a Phillips 501B Scanning Electron Microscope. The Link microanalyser was used in a qualitative mode to identify monazite ($CePO_4$), xenotime (YPO_4), zircon ($ZrSiO_4$) and uraninite (UO_2).

Whole rock XRF analyses were carried out on a Phillips PW1220 X-Ray Spectrometer in the department of Geology, University of Exeter.

Whole Rock Geochemistry

The relationship between Th, U, Zr and Ce were investigated because these trace elements appear to be strongly partitioned into the radioactive accessory mineral assemblage which, for the Carnmenellis granite, comprises monazite, apatite, xenotime, zircon and uraninite.

The variation diagram for Ce-Th is shown in Figure 2a. The lower limit of detection for Ce by the XRF technique employed is approximately 15ppm, and therefore many of the microgranite sheets, which have Ce contents below this level, are not represented.

Various authors (Alderton *et al.* 1980; Basham *et al.* 1982) have considered that a proportion of the rare earth elements (REE) are contained in the radioactive accessory minerals and do not substitute as trace concentrations in major minerals. Indeed, Alderton *et al.* (1980) in a study of REE mobility in the Cornish granites concluded that only 50% of the total REE content is present in the essential minerals; the remaining 50% being contained in zircon, apatite, sphene and allanite. However, the presence of sphene and allanite have not been detected during this study. Basham *et al.* (1982) concluded that all Ce is present in monazite, and results from this study support this conclusion.

In Figure 2a the slope of the reduced major axis for Ce/Th from whole rock XRF analyses is 4.3. This is in close agreement with the average Ce/Th ratio of monazites from Type 1 granite obtained using the Microscan 9 electron microprobe at the University of Oxford. The average Ce_2O_3 content in these monazites is 30.2% and the average ThO_3 content 8.1% giving a Ce/Th ratio of 3.6.

Figure 2b is the variation diagram for Zr - Th, for which the linear correlation coefficient is + 0.92 for the biotite granites. Since no single mineral may be invoked to explain this correlation, this implies that monazite and zircon behaved identically during granite differentiation and were depleted in the later emplaced granites. The microgranites cluster on the Zr-poor side of the linear regression line calculated from the biotite granites, implying that zircon is more highly depleted than monazite in these late differentiates.

The mean Zr content of Types I and II granites is 110ppm. Experimental Work by Watson (1979) showed that peraluminous granite magmas are saturated with respect to zircon at a level of approximately 100ppm Zr. This implies that peraluminous granites with significantly higher Zr levels contain restite zircons derived either from the source region or from partial assimilation of country rocks during the ascent of the magma. It follows from the low Zr levels in the Carnmenellis pluton that all zircons probably have a magmatic origin.

The large scatter of results on the Th-U variation diagram (Figure 2c) is presumed to be due to uranium depletion in the surface environment. Despite this scatter it appears that uranium is not depleted in the differentiation series from Type I granite through to the microgranite sheets, but remains constant at a value of between 15 and 17ppm.

An interpretation of the XRF data is that during magmatic differentiation monazite and zircon, both of magmatic origin, were depleted in the later emplaced members of the Carnmenellis pluton, whilst the concentration of uraninite remained constant.

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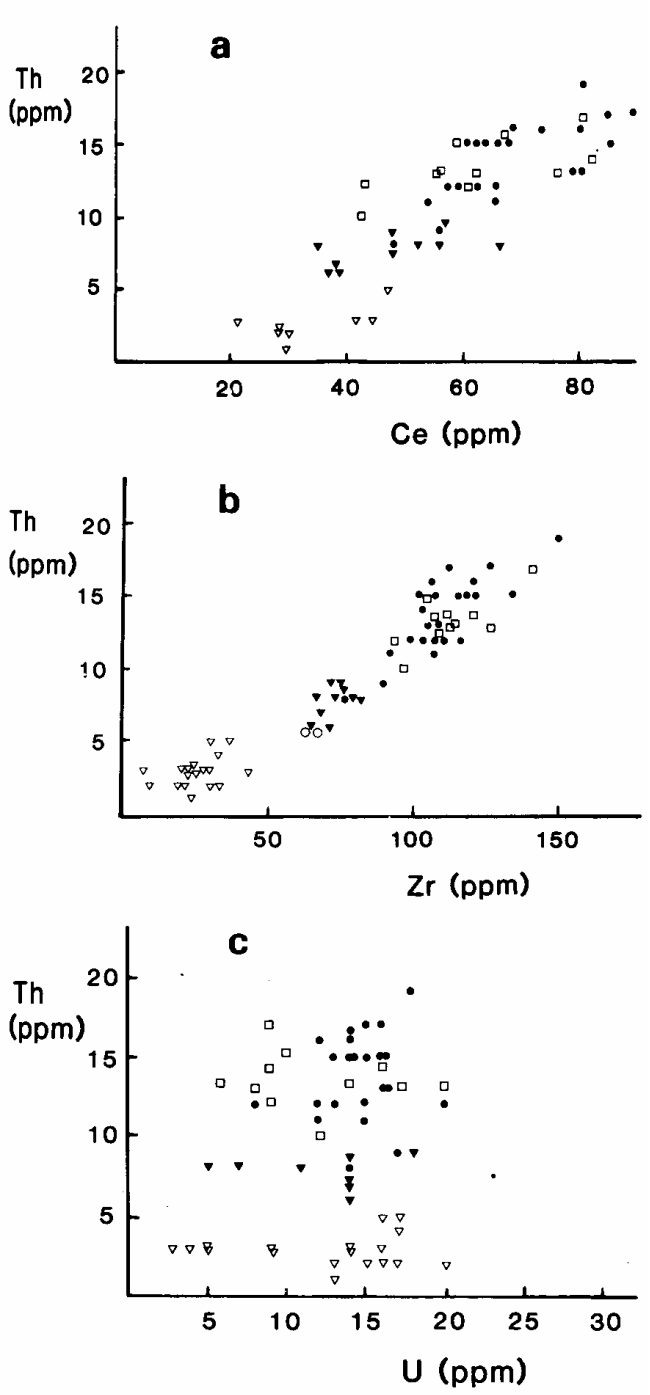

Figure 2. Trace element relationships in the Carnmenellis pluton ● Type I granite, □ Type II granite, ▼ Type III granite, ○ fine-grained granite, ▽ microgranite sheets. Data from this study and from Stone (unpublished data).

Radioactive Accessory Mineral Textures in Type 1 Granite

Stone (1979; 1984) used the 'inclusion principle' to determine the crystallisation sequence of the essential minerals within both biotite granites and lithium mica granites from south-west England. The 'inclusion principle' holds that in a crystallisation sequence the earlier mineral occurs as inclusions within the later mineral, which may totally envelop it. The present work allows a crystallisation sequence for the radioactive accessory mineral assemblage within the biotite granites to be determined, and attempts are made to relate this to crystallisation of the essential minerals.

The radioactive accessory minerals are not distributed homogeneously throughout the granite, but typically form clusters associated with ilmenite and apatite. Figure 3 shows three accessory mineral clusters, together with explanatory sketches made using the SEM and Link microanalyser. Figure 3a is an ilmenite-apatite cluster partly included in biotite and partly in perthite. From the inclusion principle this implies that the accessory mineral cluster crystallised prior to either the biotite or perthite. A similar relationship is seen in Figure 3b where an ilmenite is included in biotite, which is therefore presumed to have completed crystallisation at a later stage. Monazite and zircon occupy similar textural positions in all three clusters (Figures Sa, b, c) mostly encrusted upon the surface of ilmenite or apatite crystals. This suggests that these monazites and zircons have nucleated upon the apatite or ilmenite after crystallisation of these phases had ceased.

A general feature observed in the examination of the ilmenite-apatite clusters is that whilst both monazite and zircon are included in ilmenite, only zircon is included in apatite. An interpretation is that zircon is the first mineral to crystallise from the granite melt followed by the beginning of apatite crystallisation. It is suggested that apatite nucleated preferentially on the zircon prisms. Ilmenite must then have begun crystallising simultaneously with monazite to explain the observed inclusion of both monazite and zircon within it. The majority of the radioactive accessory minerals associated with the clusters then nucleated upon the ilmenite and apatite after these phases had ceased to crystallise. This accessory mineral assemblage is included within all of the essential minerals--biotite, plagioclase, orthoclase and quartz. This suggests that crystallisation of the accessory minerals predates the crystallisation of the essential minerals. Zircon and monazite also occur as individual crystals within all of the essential minerals, largely as randomly distributed inclusions, but occasionally marking out growth zones within the host mineral.

Uraninite is concentrated most strongly in biotite, where it forms euhedral Or subhedral crystals up to 400 pm in size. SEM photographs of uraninites separated from Type 1 granite are shown in Figure 4. Figure 4a shows a euhedral uraninite with cubo-octahedral form; striations are visible upon the octahedral faces. However, Figure 4b is more typical of uraninites separated from Type 1 granite and shows pronounced solution pitting. The solution features are considered to be largely due to groundwaters percolating through the granite in the near surface environment. As such, it is taken to be mineralogical evidence of the uranium remobilisation indicated by the XRF data. (Figure 2c).

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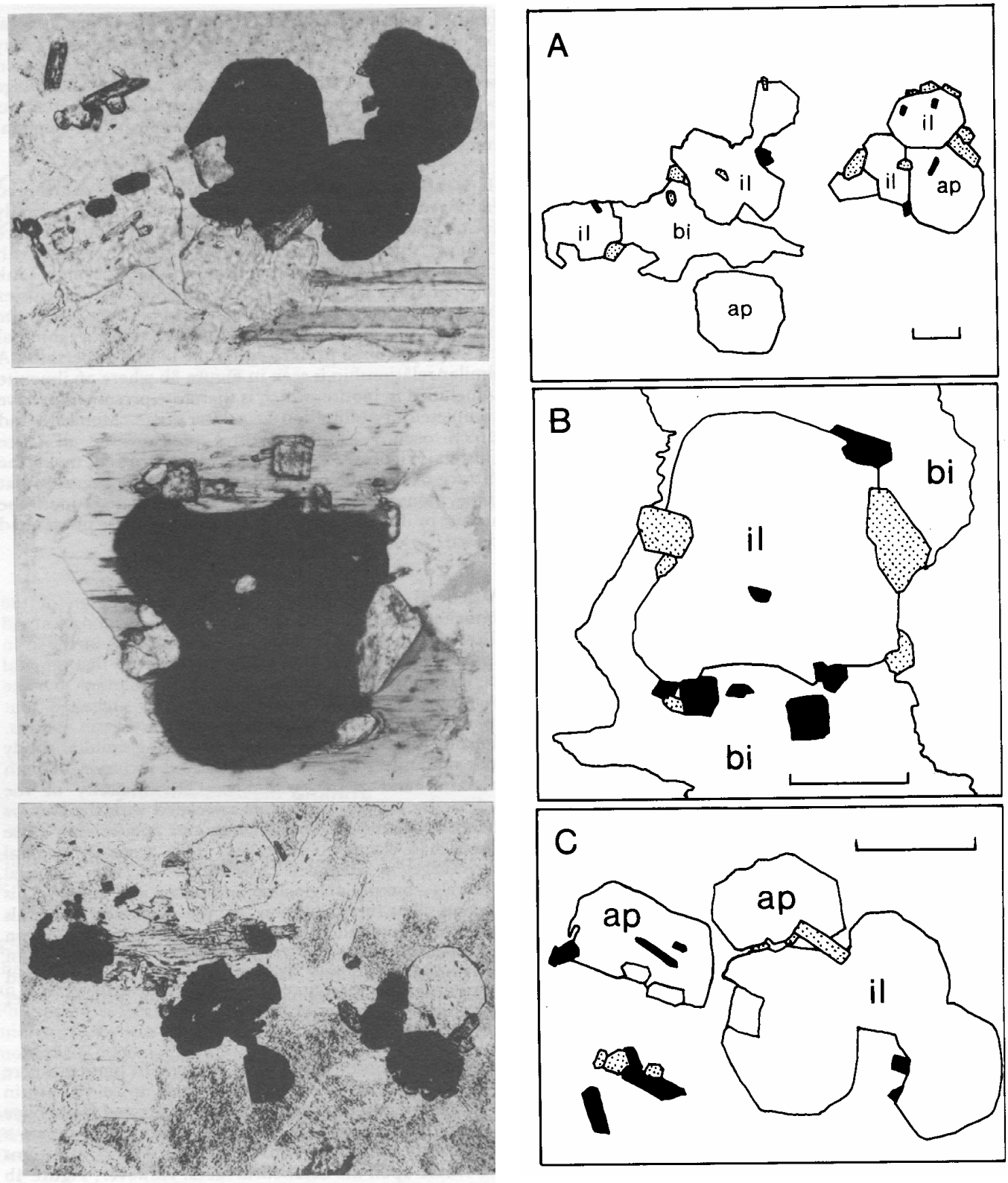

Figure 3. Photomicrographs and explanatory sketches of radioactive accessory mineral clusters from Type I granite. Zircon-black shading, monazite-heavy stipple, apatite-ap, ilmenite-il, biotite-bi. Scale bar = 100 microns.

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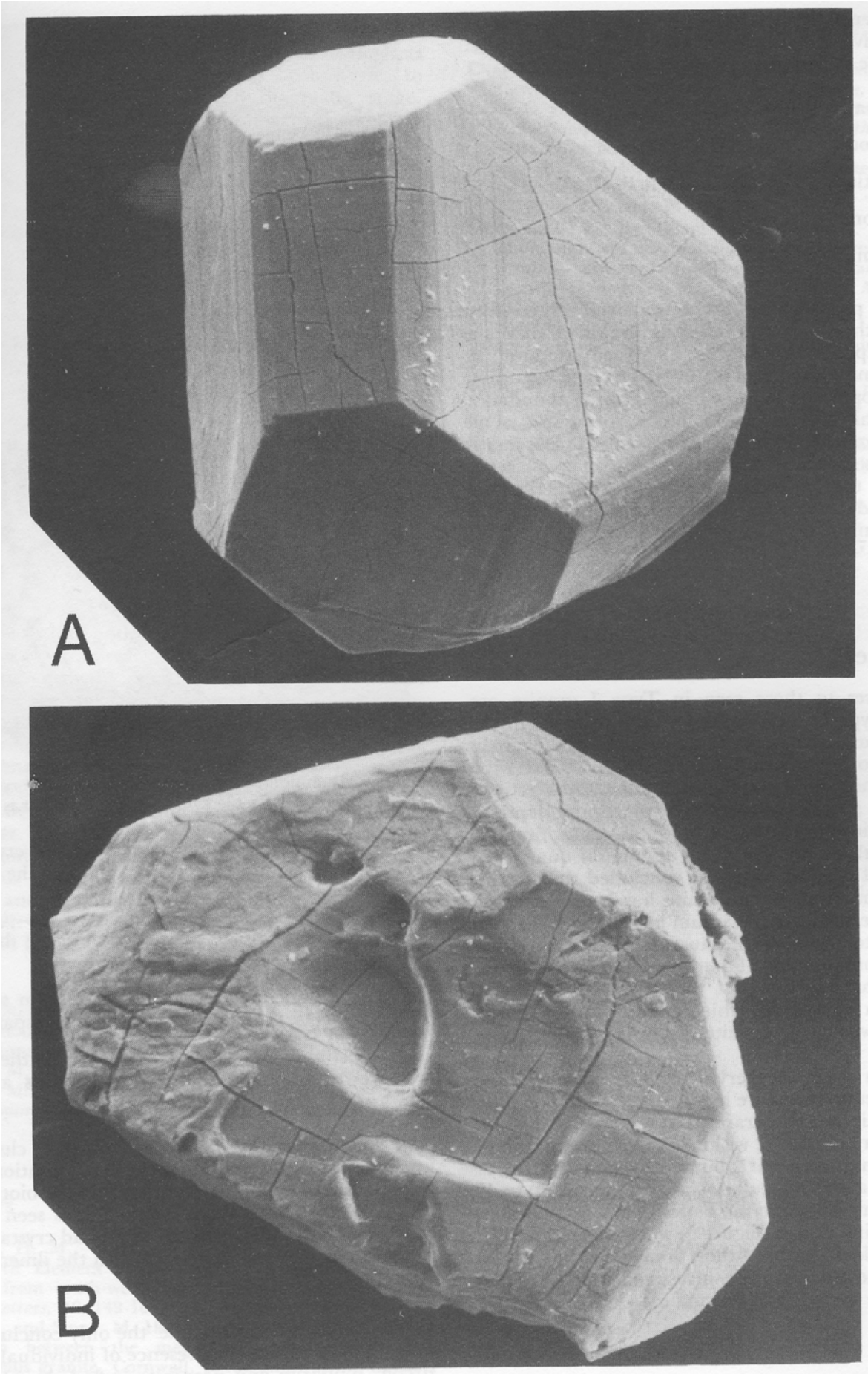

Figure 4. Uraninite crystals separated from Type I granite. (a) Euhedral crystal with a cubo-octahedral form. (b) Crystal of uraninite showing pronounced solution pitting. Depth of sample is approximately 20 metres. Field width is 300 microns.

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Table 1 presents the results of the Link-SEM study of 1200 radioactive accessory minerals from Type I granite. Five magmatic radioactive minerals were identified:

Monazite	(Ce, La, Th) PO_4
Xenotime	(Y, HREE) PO_4
Apatite	$Ca_5(PO_4)_3$ (OH, F, Cl)
Zircon	$ZrSiO_4$
Uraninite	UO_2

The relative proportions of the radioactive accessory minerals together with the minerals in which they are included, or upon which they are encrusted, are given in Table 1. The most important feature to be noted regarding the relative proportions of these minerals is the abundance of monazite which, together with zircon, forms 96% of all inclusions examined-. Xenotime is extremely rare and is only found associated with ilmenite, apatite and biotite. This suggests that xenotime crystallised simultaneously with zircon and monazite as it occupies the same textural positions within the crystallisation sequence.

Radioactive Accessory Mineral Textures in Type III Granite and Tourmaline Microgranites

Similar textures to those seen in Type I granite are observed in Type III granite, but with a decrease in the importance of the apatite-ilmenite-monazite-zircon clusters. Moreover. no uraninites were observed in Type III granite although much grain boundary alpha-activity is seen in their autoradiographs. The maximum depth of quarries within Type III granite is only 5m, in comparison with an average depth of 20m for quarries in Types I and II granite, and it is concluded that in this near surface environment uraninite has been destroyed and uranium remobilised. Uranium has not been leached from the granites, however, because the uranium content remains at approximately 15ppm. A study of the alpha-autoradiographs suggests that uranium is now largely associated with iron oxides which coat grain boundaries and fractures within the granite.

In comparison with megacrystic biotite granites, the tourmaline microgranites are highly depleted in ilmenite, monazite and zircon. Only rarely is a radioactive mineral cluster, similar to those within the biotite granites, observed. Uraninites occur abundantly as euhedral or subhedral crystals with a cubo-octahedral form, up to 40 µm in diameter. They are most prominent as inclusions within tourmaline prisms (Figure 5) where they form blue pleochroic haloes within the orange/brown schorl crystals. In one microgranite investigated 85% of the uraninites were included in tourmaline.

Discussion

From an examination of the textures of the biotite granites Stone (1979) concluded that several distinct parts of the granite fabric could be recognised, and that these

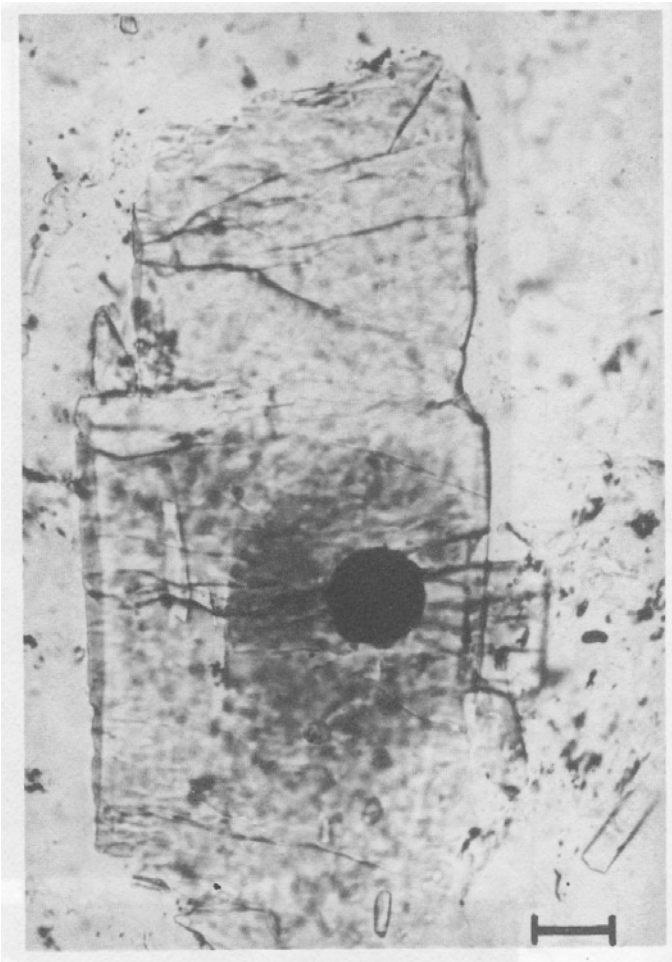

Figure 5. Inclusion of uraninite within a schorl crystal, showing the development of a pleochroic halo. Scale bar is 25 microns.

can be related to different stages in the crystallisation sequence. Using the 'inclusion principle' he recognised the following stages:

(1) a largely restite phase comprising biotite and andalusite

(2) a magmatic stage with a cystallisation sequence of plagioclase feldspar-quartz-alkali feldspar, and

(3) a postmagmatic stage represented by the continued growth of alkali feldspar megacrysts and by the growth of tourmaline and muscovite.

The occurrence of accessory mineral clusters with magmatic textures predating the crystallisation of biotite implies that much of the biotite within the biotite granites must be magmatic. From Table ! it is seen that more than half of all radioactive minerals had crystallised prior to biotite as they are associated with the ilmenite-apatite clusters.

Using the 'inclusion principle' the only conclusion which may be drawn from the presence of individual crystals of zircon, monazite and uraninite included within biotite, feldspars and quartz is that the essential minerals have continued to crystallise after nucleation of the accessory minerals. However, these accessory minerals also occur at the edges of both biotite and plagioclase crystals. This

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texture is identical to the 'encrusting' of ilmenite and apatite by zircon and monazite, and therefore suggests that radioactive accessory minerals continued to crystallise towards the end of the magmatic stage. Approximately one third of all radioactive accessory minerals occur as individual crystals within biotite whilst less than 12% form individual crystals (that is, not associated with ilmenite-apatite clusters) within the feldspars and quartz (Table 1). The implication is that most radioactive accessory minerals had crystallised prior to, or simultaneously with, biotite and that only 12% nucleated during crystallisation of the felsic minerals.

Table 1. The Percentage Distribution of Radioactive Accessory Minerals in Type 1 Granite, Carnmellis

	Radioactive Mineral Inclusion				
Host Mineral	Zircon	Monazite	Uraninite	Xenotime	Total
Ilmenite/Apatite	31	22	0.6	0.8	54.4
Biotite	12	20	1.5	0.1	33.6
Plagioclase	2.0	4.0	0.5	-	6.5
Quartz	2.0	1.5	0.2	-	3.7
Perthite	0.6	1.0	0.2	-	1.8
Total	47.6	48.5	3.0	0.9	

Tourmalines within both the biotite granites and the lithium mica granites have been regarded as postmagmatic growths (Stone, 1979; 1984). Both prismatic and skeletal crystals were observed by him in the lithium mica granites and the former was interpreted as having been precipitated directly from a fluid/vapour phase. Tourmaline crystals within the microgranite sheets are prismatic (Figure 5) with no evidence of interfingering of the type described by Stone (1984)from the lithium mica granites. The implication is that tourmaline within the microgranites has precipitated directly from a boron-rich aqueous fluid. From the close relationship of tourmaline and uraninite within the microgranites this further implies that uranium is being transported in a aqueous fluid in the last stages of granite differentiation.

Acknowledgements I wish to thank C. J. Brice; J. R. Merefield and Miss N. Hooper for carrying out the trace element analyses, and Dr. M. Stone for allowing me to use some of his unpublished data. Dr. E. M. Durrance kindly read an earlier draft of this paper, and Mrs. C. Holmes typed the final manuscript. Financial support for this study was provided by the NERC and CEGB.

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Read at the Annual Conference of the Ussher Society, January 1984

Emplacementof the Porthmeor granite pluton, West Cornwall

M. STONE
C. S. EXLEY

Stone, M. and Exley, C. S, 1984. Emplacement of the Porthmeor granite pluton, West Cornwall, *Proceedings of the Ussher Society 6*, 42-45.

A small biotite granite pluton exposed in Porthmeor Cove, near Zennor, Cornwall, was intruded into Mylor metasedimentary rocks and metadolerite by joint-controlled passive emplacement, and subsequently differentiated *in situ*. Evidence that the pluton post-dates the emplacement of the main Land's End granite is provided by its vertical contact which cuts across structures in the envelope that had previously been tilted by the Land's End granite and by cross-cutting relations among associated granite dykes.

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Introduction

Porthmeor Cove (SW425375) is situated about 3 km WSW of Zennor on the NW coast of the Penwith peninsula. Two small granite intrusions, with associated veins and dykes, are exposed on the eastern side of the cove, the more northerly, however, being in the cliff and inaccessible. The other, which is the subject of this account, is well displayed in three dimensions in the rocks near the base of ttie cliff just below Long Carn (Fig. 1).

Despite being very well known to many geologists, few references to the intrusion seem to have been published, those of which we are aware being limited to Reid and Flett (1907), Booth (1966), Hall (lg74), Hall and Jackso (1975) and Exley and Stone (1982). All these are extremely brief and since the rocks show a number of features of interest and significance we offer here some more detailed information and a chronological interpretation of the igneous events.

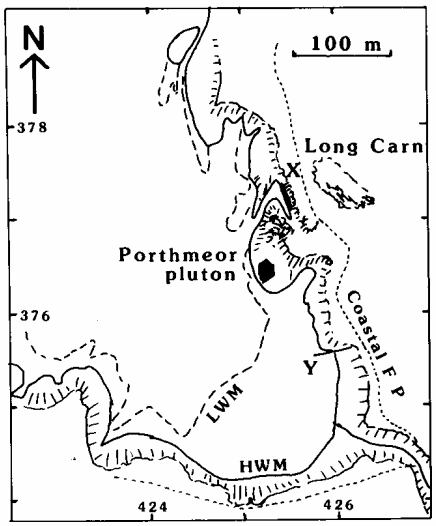

Figure 1. Sketch map of Porthmeor Cove showing the position of the Porthmeor pluton (black), the more northerly inaccessible granite dome (X) and the main Land's End granite contact (Y), with metadolerite to the north overlying granite to the south. The contact dips about 20 degrees to the north.

Field relations

The country rocks are banded Mylor pelitic hornfelses which are underlain by a massive metadolerite sheet. The general dip of the bedding in the pelites above the metadolerite is about 20 degrees to the north. Since the junction between the pelites and the metadolerite dips at the same angle and the latter is devoid of volcanic characteristics, it is presumably a sill. Higher in the cliff above the pluton, the axial planes of F3 folds (Smith, 1962; Exley and Stone, 1982) in the pelites also have a northerly dip close to 20 degrees.

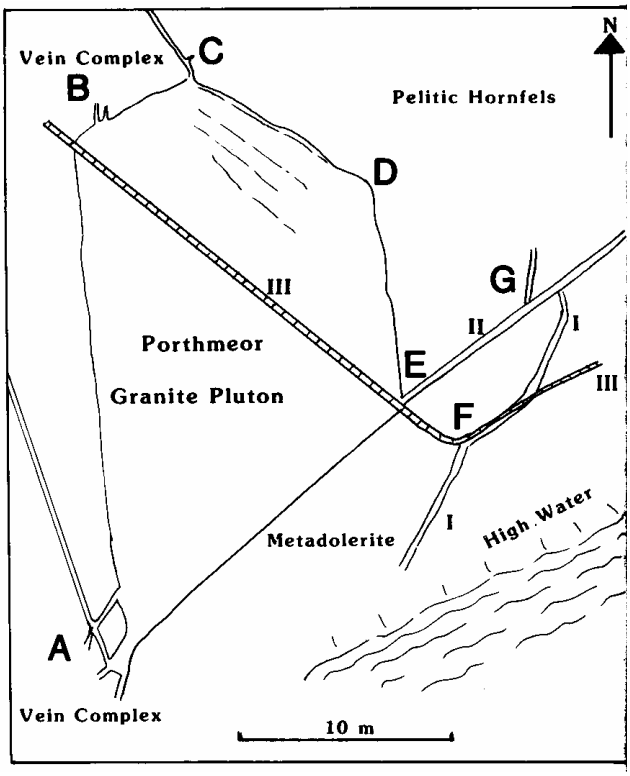

Figure 2. Sketch map of the Porthmeor granite pluton showing dykes I, II and III and localities A - G described in the text.

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The intrusion constitutes a small boss, parts of whose roof and near vertical walls are exposed. In plan it is irregularly pentagonal, measuring some 19 by 15 m, with sharply angular corners and in detail the walls are also irregular with angular re-entrants (Fig. 2; sections AB, BC, and AE). Clearly, emplacement has been controlled by the joint system in the surrounding rocks, the principal directions being NE-SW and NNW-SSE. The roof is gently domed and all contacts are sharp. A cross-section based upon colour slides taken with a telephoto lens from the west side of the Cove is shown in Figure 3 and compares quite closely with the diagrammatic section given in Exley and Stone (1982; Fig. 21.3).

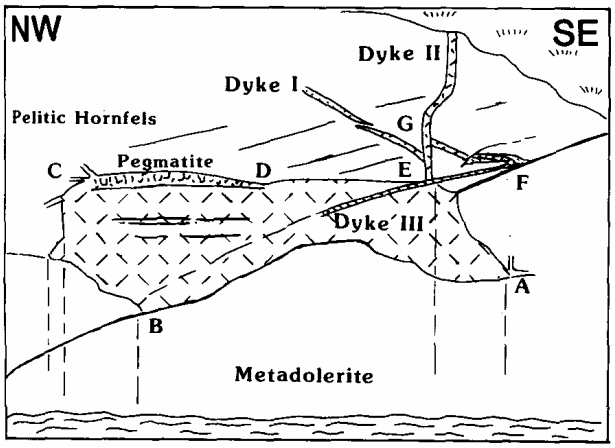

Figure 3. The Porthmeor pluton as seen in section viewed from the western side of Porthmeor Cove. The pluton is shown by diagonal dashed lines and with roof pegmatite at its northern end. The three dykes I, 11 and III and the lettered localities A-G are shown in Figure 2 and referred to in the text. Scale—the distance CE is about 18 m. The line that continues BF outside the granite is the junction between metadolerite (below) and pelitic hornfels (above).

Angular networks of granitic veins that clearly originate from the pluton occur at corner A and between corners B and G (Fig. 2).

The pluton is composed of megacrystic biotite granite (Type B; Exley and Stone, 1982). However, this shows some variation in texture, especially towards the middle where distinct layers of darker coloured aphyric granite result from concentrations of biotite. Beneath the layered zone, the granite is sparsely megacrystic, but again shows some textural variation and may prove on detailed analysis to be inhomogeneons. Beneath the roof contact, there is a development of a narrow, but typical pegrnatite/leucogranite complex about 60-70 cm thick. This has developed under a slight dome or rise in the roof in the northern half of the section. It is similar to the layered complex that occurs in the roof zone of the Tregonning granite and in its associated granitic sheets at the Megiliggar Rocks (Stone, 1975). As in the Tregonning granite, it seems likely that the pegmatite at Porthmeor has deve1oped beneath a structural trap where volatiles accumulated as the granite was finally solidifying from magma. By analogy with the Tregonning granite, this, together with the horizontal layering and lack of evidence for the later emplacement of any of the layers, indicates differentiation *in situ.*

Dyke phases

Figure 2 shows a number of granitic dykes associated with the Porthmeor pluton. Those numbered I, II and III are the most significant in terms of chronology. Dyke II is about 50 cm wide and appears, like the vein complexes at A and near B in Figure 2, to arise directly from the pluton. It is comp. osed of megacrystic granite, and in hand specimen is petrographically similar to the megacrystic granite in the main body of the pluton. This dyke, which is almost straight for most of its length, cuts and markedly offsets Dyke I about 8 m from the junction of Dyke II with the pluton (at G in Figs. 2, 3 and 4). Dyke I, about 30 cm thick, is composed of leucogranite. Dyke III, which is of tourmaline microgranite, is only some 12 cm thick where it meets and enters Dyke I. After entering Dyke I, Dyke III gradually transgresses it obliquely in a south-westerly direction as far as point F, whence it suddenly changes to a north-westerly direction and leaves Dyke I. The relationships between these dykes are shown in the synoptic diagram of Figure 4. This is based upon colour slides that show each stage of the relationship from point G, through point F to point E in Figure 2. After changing direction and leaving Dyke I, Dyke III crosses Dyke II where the latter joins the pluton and then cuts right across the main body of the pluton in a deep gully. Dyke I forms the steep rock wall of the gully to the south of the pluton and joins the underlying layered complex apparently without discordance.

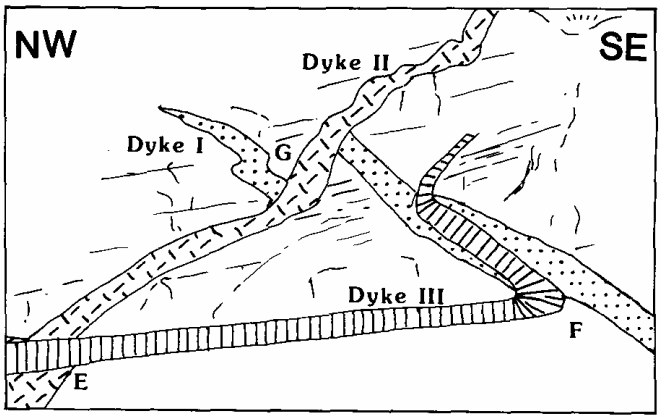

Figure 4. Synoptic drawing showing relations between dykes I, II and III, based upon colour slides. Localities E, F and G as in map, Figure 2. Scale—perspective drawing: bottom width about 5m; horizontal width through G about 20 m.

Relationship to main Land's End granite

The contact between the main Land's End granite and its country rocks is exposed about 100 m south of the Porthmeor pluton and here strikes roughly E-W, dipping about 20 degrees N under massive metadolerite (Figs. 1 and 5). This is also Type B granite and has a layered roof complex of leucogranite, aplite and pegrnatite that overlies a coarsely megacrystic texture so typical of these granites near roof contacts.

There is also a layered leucogranite/aplite zone underlain

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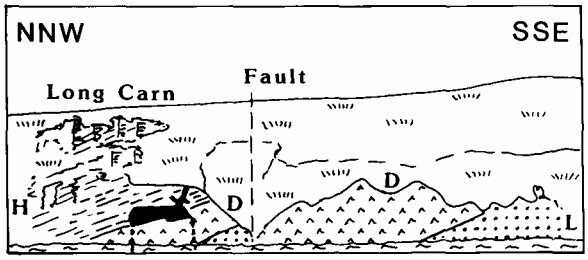

Figure 5. Section along eastern side of Porthmeor Cove, drawn from photographs. The top of the Porthmeor pluton is shown in black and dotted lines extend it downwards behind the meta-dolerite (above P). D (pecks) — metadolerite; H (dashed lines)— pelitic hornfels; L (dots) —main Land's End granite and layered granite exposed beneath the Porthmeor pluton at low water. The width of the section is about 275 m; that of the pluton in this section is about 25 m.

by megacrystic biotite granite visible at low water immediately below colder A of the pluton (Fig. 2). These rocks may be either part of the main granite, with a contact still dipping about 20 degrees N, uplifted by a fault which is visible in the adjacent gully in the cliff (Fig. 5), or the lower part of the Porthmeor pluton itself; The layering is similar to that near the roof of the pluton and the contact is stepped like that seen at Rinsey (SW593269) in the Tregonning granite, though on a small scale. However, the dyke relationships described above suggest that this lower body of layered granite predates the emplacement of the pluton, since Dyke I appears to arise from it and is cut by dyke II, which appears to be part of the pluton.

Emplacement

From the relations described above, it seems that the order of emplacement is:

1. Dyke I and the lower layered complex.
2. The Porthmeor pluton, Dyke II and the vein complexes.
3. Dyke III.

Since the Porthmeor pluton cuts vertically through the inclined Mylor metasedimentary rocks and associated metadolerite without disturbing any of the contacts, it must be inferred that the main Land's End granite had been emplaced and had tilted its envelope some 15-25 degrees prior to its emplacement. Suporting evidence is provided by the lower layered granite from which Dyke I evidently originated and which could be part of the roof complex of the main Land's End granite. Like the main Land's End granite in the Cove (Fig. 1), this lower granite has layering conformable with its contact and with the dip of the nearby country rocks.

The lack of disturbance of the country rocks, together with the joint-controlled contacts, indicates that the mechanism of emplacement was essentially passive, involving 'block' or, more likely, 'cauldron' subsidence on a small scale. Confirmation of this interpretation is suggested by the presence of angular xenoliths stoped from the roof in the more northerly plutort in the cliffs, and by the nature of the stepped contacts of the Rinsey roof pendant, near Porthleven, in the structurally similar Tregonning granite.

Conclusions

It is concluded that the Porthmeor pluton provides another, though minor, example of the passive emplacement of granite magma, by stoping or cauldron subsidence. The emplacement was clearly joint-controlled and was followed by differentiation *in situ.* The production of pegmatite in a slight dome in part of the roof was the result of extreme differentiation and the concentration of the small residual magmatic volatile fraction. In these features, it resembles the Tregonning granite.

As a general feature, the country rock envelope around the Land's End pluton dips away from the granite. At Porthmeor, this tilted envelope of pelitic hornfels underlain by metadolerite has been cut vertically by the Porthmeor pluton. The tilting of the envelope around the Land's End granite must have occurred either at the time of its emplacement or in the interval between this event and the emplacement of the Porthmeor pluton and its associated dykes. A post-emplacement tilting of the envelope could be accounted for by uplift resulting from isostatic readjustment (Bott *et al.*, 1958), but such adjustment would have occurred before the intrusion of the Porthmeor pluton. This seems unlikely since the plutonic events probably followed one another without any significant break.

Thus, whilst the dominant mechanism of emplacement of the plutons belonging to the Cornubian batholith was subsidence and stoping, a small diapiric component was present in the case of the Land's End pluton. This produced tilting of formerly near-horizontal structures in the envelope, before the Porthmeor pluton was emplaced into both underlying granite and its tilted envelope.

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Read at the Annual Conference of the Ussher Society, January 1984

Xenolith assimilation in the granites of south-west England

C. J. LISTER*

Lister, C.J. 1984. Xenolith assimilation in the granites of south-west England. *Proceedings of the Ussher Society, 6,* 46-53.

The granites of south-west England show unusual enrichment in elements such as Sn and W, which are virtually absent from many other granites of similar general composition. Xenolith assimilation is assessed as a possible mechanism for introducing additional elements to the granite magma, with particular emphasis on those elements that are important in mineralising processes.

Geochemical gradients from sediments through aureole rocks and xenoliths to granites indicate that Sn, W, U and Ta cannot have been brought from sediment into granite by the assimilation process, while V, Ba, Sr and possibly Cu and Zn could be derived by this route. Li, F, and Th are probably derived from the granite, but are preferentially concentrated in xenoliths, which may provide a trap for these elements in granite roof zones by virtue of their high biotite concentrations.

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Introduction

The Variscan granites of south-west England are emplaced within low grade metasediments of Devonian and Carboniferous age, and contain abundant xenoliths representing included fragments of country rock. Xenoliths are particularly common in the megacrystic granite facies of Dartmoor and the St. Austell and Land's End plutons. Most xenoliths resemble the biotite hornfels found at granite contacts, and can reasonably be regarded as metamorphosed pelitic sediment.

The granites conform with the aluminous 'S-type' of Pitcher (1982) on the basis of their geochemical and isotopic compositions, their associated mineralisation, and their xenolith assemblages. According to Hampton and Taylor (1983), Pb and Sr isotopes indicate a crustal source in Proterozoic basement about 800 m.a. old. The present composition of the granites must be dependent firstly upon the nature of this basement protolith, and secondly upon processes of upper crustal assimilation, as well as upon the action of post-magmatic fluids. The unusual enrichment in certain elements (Sn, W, U, Li, F, B) found in this province cannot be explained simply by the nature of the inferred protolith, since other, unmineralised granites (e.g. Mountsorrel (Hampton and Taylor, 1983)) are apparently derived from a protolith with a very similar age and Pb and Sr isotopic characteristics. Hence, some external source for these elements must be sought, either at the stage of magma generation, or in the later history of the granites. A mantle contribution to mineralisation cannot be ruled out by the available data, but the present aim is to consider the possible role of upper crustal contamination.

Xenoliths provide a clue to the degree of contamination affecting the granites, and should yield some evidence to show whether assimilation in the upper crust is a feasible mechanism for the supply of individual elements. Assimilation followed by differentiation could account for the availability of residual fluids rich in metals, as well as fluorine and boron, whose effects are relatively localised, but it is also worth considering the extent to which upper crustal material could have influenced the general geochemical character of the granites. The possibility that an entire pluton could be generated from upper crustal sediments was examined by Tindle and Pearce (1983) who found that greywacke xenoliths in the Loch Doon granite, Scotland, could be equated with a trondhjemitic composition. These authors remarked that no major intrusions of this composition were known; it has, however, been suggested that the south-west England granites were originally trondhjemites, and that their present composition is due to subsolidus effects (Hawkes *In* Edmonds *et al.*, 1968).

Brammall and Harwood (1932) realised the importance of xenoliths in the Dartmoor granite, and noted the exchange of alkalis and alumina across granite/xenolith boundaries. In addition to analysing several xenoliths for major and a few trace elements, these authors made a geochemical traverse across a pelitic xenolith, and found Al_2O_3Ancreasing outwards to a maximum at the margin reflected by a concentration of muscovite and tourmaline; they decided that since MgO, CaO, TiO_2, alkalis, and total Fe oxides showed decreasing concentrations towards the xenolith margin, these components were effectively being added to the magma.

The purpose of the present investigation has been to identify trace elements that might have been added to the magma by assimilation, with particular emphasis placed on those elements that are important in post-magmatic mineralisation processes. A study confined to granites and xenoliths only could give misleading results, since the transfer of ions between granite and xenolith could enrich the xenolith in elements derived from the granite (a

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supposition supported by the results of this work). Therefore, where possible, contact metasediments within the granite aureole were also sampled, as well as sediments away from the thermal effects of the granite and subjected to low grade regional metamorphism only. The latter will be referred to as 'sediments' in the following pages, in order to distinguish them from 'metasediments' of the aureoles. The provenance of samples is indicated in Figures 1-3.

Distribution of Xenoliths

The greatest abundances of xenoliths in south-west England are found in megacrystic granite where it occurs on Dartmoor, the eastern St. Austell mass, and around the Land's End coastline, although xenolith concentrations are also present in aplite/pegmatite roof complexes (e.g. Megiliggar Rocks, Exley and Stone, 1982). The granite facies now known as the small megacryst variant (Dangerfield and Hawkes, 1981), typified by the Bodmin Moor granite, contains relatively few xenoliths but shows evidence of contamination in the form of streaked-out biotite-rich concentrations which may represent assimilated xenoliths. The later intruded, fine-grained granites appear to contain no xenoliths.

There is certainly a concentration of xenoliths close to the granite roof, and this could be taken to indicate either that granite/sediment interaction only affects the upper levels of plutons, or that at depth all xenoliths are completely assimilated by the granite. In view of the relict xenolith textures in some granites at a distance from the contact, the latter interpretation seems more likely; the presence of andalusite, cordierite and other sedimentary derived xenocrysts within parts of the Dartmoor granite (Brammall and Harwood, 1932) also supports this view.

Pelitic xenoliths range in size from over 1 m to less than 1 cm, and vary in shape from angular fragments to sub-spherical bodies. The degree of assimilation also varies, so mat xenolith composition may be anywhere between biotite hornfels and a rock virtually indistinguishable from granite.

Examination of different granite contacts in south-west England reveals that contacts are not simple, straight lines, but irregular junctions involving apophyses and veins that detach roof fragments along joints and cleavages. A few examples will illustrate the nature of interaction between sediment and granite at such contacts.

At Megiliggar Rocks (SW612266), the contact between aplite sheets of the Tregonning granite roof complex and spotted slates is marked by pegmatitic development with crystals growing vertically downwards into the aplite (Stone, 1969); xenoliths within the aplite are generally very angular and tend to act as volatile traps with pegmatitic bands below. The adjacent bay, Rinsey Cove (SW593269) 'is eroded into the well-known 'roof pendant' (Stone, 1975), with metasediments forming the flat floor of the cove, and granite outcropping on three sides; the granite at the western contact terminates in pegmatite and large angular xenoliths are preserved below this.

At Cape Cornwall (SW354321), the contact between megacrystic granite and metasediment on Porthledden beach follows a relatively straight line between high and low water mark. There are complex relationships between two granite types (megacrystic and fine-grained) and a large tourmalinite body (Charoy, 1979; Lister, 1981), but the megacrystic granite forms a discrete unit in which xenoliths are abundant. The fine-grained granite contains no xenoliths. Xenoliths in the megacrystic facies are generally rounded in shape. A finely-crystalline biotite-rich type with feldspar megacrysts is quite common, but both aphyric and very coarse-grained xenoliths are also present; it is possible that the xenoliths are polygenetic, since although the contact metasediment is a spotted hornfels, andalusite-bearing hornfels and metabasites also occur in the vicinity.

On Leusdon Common, East Dartmoor, a horizontal contact between megacrystic granite and spotted hornfels is exposed in a small outcrop (SX704728). Several granitic veins cut across the hornfels, isolating angular roof fragments. Large granite boulders nearby contain numerous xenoliths in various stages of assimilation. Some boulders consist of intimate mixtures of granite and xenolithic material with alternating leucocratic and biotite-rich bands, and textures resembling those of migmatites. Clearly the degree of contamination of the granite at this locality is considerable.

Certain areas well within the major plutons also furnish examples of xenoliths in various stages of assimilation, and these are probably exposing granite close to the original roof, e.g. Chagford, north-east Dartmoor (SX705867), and Luxulyan in the St. Austell granite (SX054590). Although Dartmoor was unroofed as early as the Permian, there is evidence to suggest that the present erosional level is not greatly different from that immediately following the original unroofing (Dangerfield and Hawkes, 1969).

The material for the present investigation came from Dartmoor, the St. Austell granite, and the Land's End granite, with their surrounding sedimentary envelopes. The Bodmin and Carnmenellis granites are not represented for the simple reason that no xenoliths suitable for chemical analysis were found within these granites. The Tregonning-Godolphin granite is not represented because xenoliths found were in roof complexes rather than in true granite.

Progressive Assimilation of Pelitic Sediment

The country rocks in contact with the south-west England granites are predominantly metapelites and metasandstones, with local occurrences of calcareous metasediments and metabasites.

The present work has concentrated on metapelites and pelitic xenoliths for two reasons. The most common metasedimentary lithologies of this province are likely to represent the greatest volumetric contribution of assimilated material, and the metapelites are richer in trace elements than the metasandstones and are therefore more likely to have a recognisable geochemical signature.

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A series of stages in the progressive assimilation of pelitic sediment can be identified, starting with sediments outside the metamorphic aureole of the granites, passing through aureole rocks, then through xenoliths, and finishing with megacrystic granites that contain assimilated xenoliths. In the present investigation, samples were collected to represent the following stages in this process:

1. Pelitic sediments outside the metamorphic aureole These include Devonian grey shale or slate consisting of clay minerals ± chlorite and white micas, and finely divided quartz which is sometimes concentrated in quartz-rich horizons, as well as Carboniferous black shales, rich in organic material, and Carboniferous calcareous shales.

2. Aureole metasediments. The most commonly encountered lithology is a spotted slate with spots of pinitised cordierite, anhedral but regularly aligned biotite, quartz and opaque oxides. Preferred orientation of mineral grains is prominent.

3. Metasediments at the contact with granite. These may be spotted slate, spotted hornfels or biotite hornfels. In some cases, a preferred orientation of minerals is preserved right up to the contact (perhaps indicating relatively low temperature of the granite or aplite at the time of emplacement), while in others, the contact rock is a hard hornfels, lacking any preferred orientation. Well crystallised biotite is ubiquitous: other minerals encountered include pinite, tourmaline with colour zoning, quartz and opaque oxides.

4. Xenoliths. All xenoliths examined contain biotite as the predominant mafic mineral; they can be categorised according to the presence or absence of foliation or feldspar, with gradations between categories.

a) *Foliated xenoliths, retaining a metamorphic preferred orientation.* These xenoliths appear to have undergone relatively little recrystallisation during residence in the granite, and can therefore be taken to represent the earliest preserved stage of assimilation: they are not found at all xenolith localities since in many cases the foliation has already been lost from the metasediments at the contact. These xenoliths consist almost entirely of oriented biotite and quartz with opaque oxides, ± tourmaline. Radiogenic inclusions in biotite are common; at this very early stage in the assimilation process, radiogenic material is evidently available to become incorporated in zircons (and possibly in other accessories) within the recrystallising biotite.

b) *Unfoliated xenoliths.* Even within less than '1 m of the contact, many xenoliths lack foliation and have a fine-grained and even texture. This is generally (but not always) true where the contact metasediments are hornfelsed and it may be presumed that, by comparison with foliated Xenoliths, a greater: degree of recrystallisation has occurred, either prior to detachment from the granite roof, or during residence in the granite. Since they have essentially undergone a higher grade of metamorphism and more closely approach the granitic texture, these xenoliths can be considered to represent a slightly more advanced stage of assimilation than the foliated variety; the mineralogy of the two types is very similar, including the presence of biotite with radiogenic haloes.

c) *Feldspar-bearing xenoliths.* This category includes all xenoliths that have a similar mineralogy to the enclosing granite, although the proportions of components may differ. As assimilation proceeds, biotite recrystallises to larger grains in a more random orientation, retaining haloes. Alkali feldspar megacrysts similar to those in the granite appear within xenoliths that still retain a high biotite concentration; some xenoliths at this stage also contain tourmaline replacement pods with leucocratic margins. Further recrystallisation increases grain size and reduces the biotite concentration, until eventually the xenoliths are almost indistinguishable from the granite. In advanced stages of assimilation, those xenoliths would certainly be regarded as granite if seen alone. With this fact in mind, it is perhaps not surprising that many granites giving no indication of hydrothermal alteration nevertheless show apparent replacement textures; within totally assimilated xenoliths, every mineral grain is effectively a secondary replacement product.

5. Granite. The granite samples used in this work are all megacrystic varieties that contain abundant xenoliths. The granite comprises perthitic alkali feldspar megacrysts biotite and quartz, .+sodic paligoclase, -+muscovite, + tourmaline, with accessory zircon, apatite, rutile and/or ilmenite.

An ideal progressive assimilation sequence would include all the above rock types from a single locality or a single granite contact. The nearest approach to this ideal was achieved in the area of Merrivale, West Dartmoor (SX547755), where all members of the sequence were obtained except the metasediment at the actual contact, and at Cape Cornwall which yielded all except the sediment outside the metamorphic aureole (this presumably occurs offshore). Only at Luxulyan and Chagford was it possible to examine xenoliths of apparently the same parent rock in different stages of assimilation; otherwise all xenoliths from the same locality were considered together. More fragmentary data were obtained from Sennen Cove (SW348263), Corndon Tot (SX686742), below Leigh Tor (SX712714) and Burrator Reservoir Quarry (SX550678).

Geochemical Gradients in Progressive Assimilation

Major and trace elements were determined by inductively coupled plasma emission spectrometry (ICP) at King's College, London, with the exception of Sn (ICP at Imperial College), F (selective ion electrode), and W, Ta, U, Th, Ta and rare earth elements (REE) (neutron activation analysis at Imperial College Reactor Centre). The data for selected trace elements are presented in Figs. 1 to 3, and the distribution of elements is summarised in Table 1.

If a particular element were introduced to the granite by

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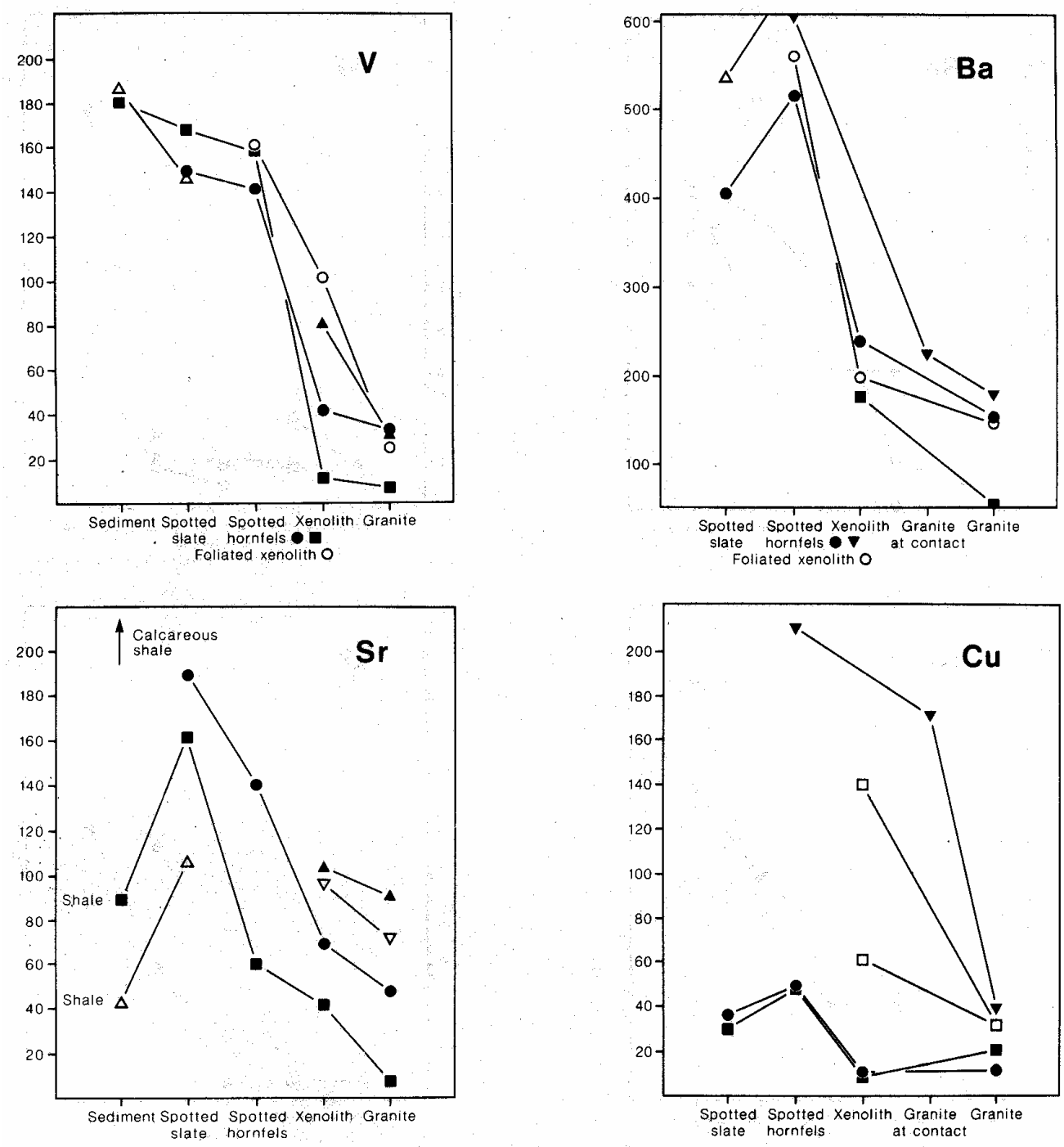

Figure 1. Geochemical gradients for vanadium, barium, strontium and copper. All values ppm.
Legend: ■ Merrivale ● Cape Cornwall ▲ Sennen Cove ○ Luxulyan □ Chagford △ Leigh Tor ▽ Corndon Tor ▼ Burrator

the process of assimilation, then a continuous geochemical gradient might be expected, from relative enrichment in the sediments and metasediments, through an intermediate value in the xenoliths, to lower values in the granites. Elements that show such a trend could be derived from sediment rather than granite, although the degree of transfer would depend upon partition coefficients of individual elements between xenolith and granite, or between xenolith and fluid. This trend is shown by V, Ba, Sr and Cu (Fig. 1). The most pronounced concentration gradient is shown by vanadium, which is highly enriched in black shales, and progressively less enriched in metasediments, xenoliths and granites. It appears that vanadium is fixed in the metamorphic minerals of some metasediments at the contact with granite, where vanadiferous biotite and tourmaline were identified by electron microprobe analysis. Barium shows a similar trend to vanadium. Although the highest concentrations of barium are recorded in metasediments at the contact with granite, sedimentary concentrations are nearly always higher than those of either xenoliths or granites; a derivation of Ba from shales was in fact suggested by Brammall and Harwood (1932). The concentration of strontium in

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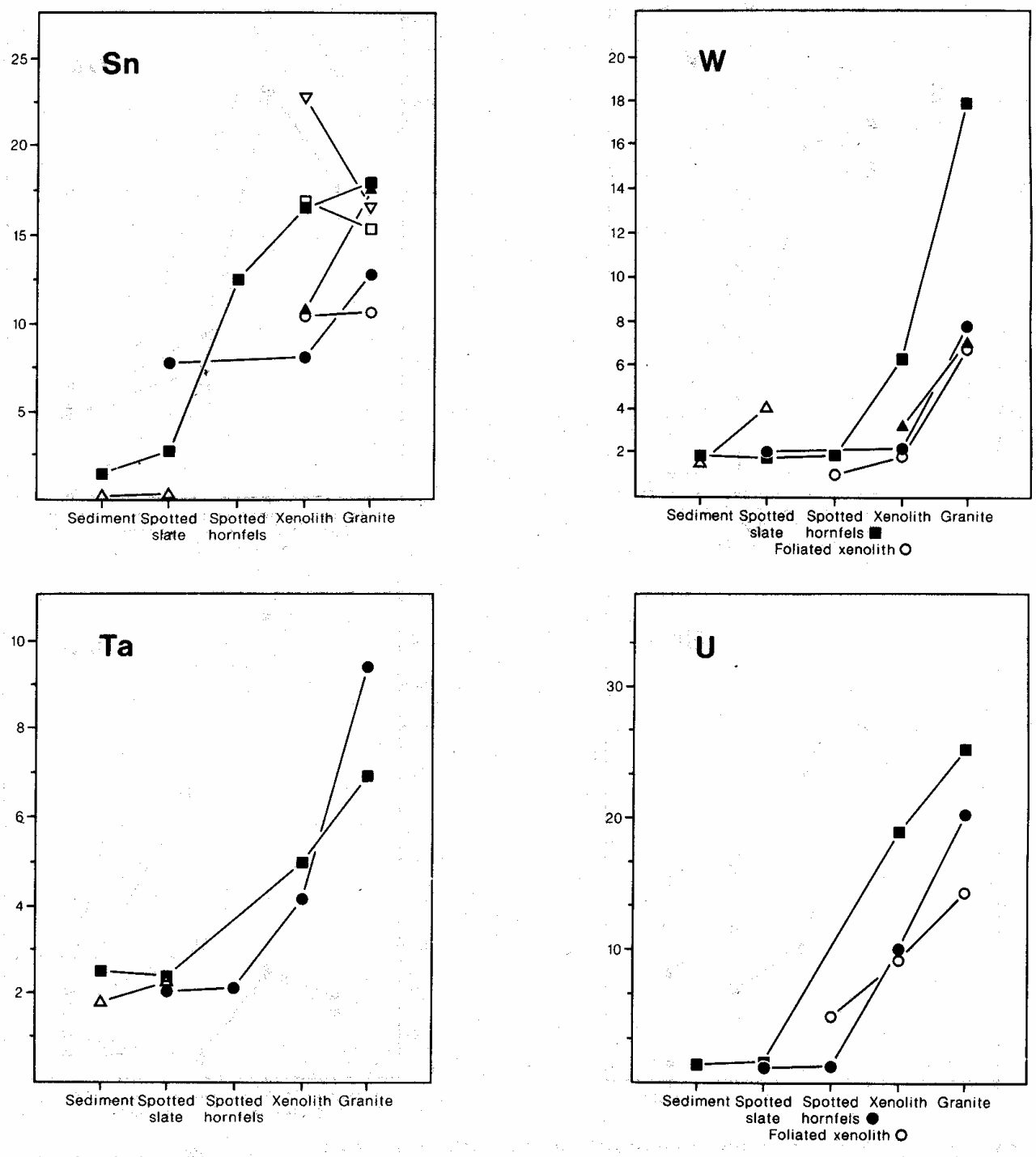

gure 2. Geochemical gradients for tin, tungsten, tantalum and uranium. All values ppm. Legend as for Figure 1.

sediments is highly variable, following calcium, so that in calcareous shales, Sr reaches over 1000 ppm. In xenoliths and granites, Sr values are low, and the general concentration gradient is similar to that of barium. Copper also shows relatively low concentrations in granites, but is highly enriched in some metasediments in contact with granite; sedimentary concentrations are frequently higher than those of the granites (the same is true of Zn at some of the localities sampled), it is worth noting that Floyd (1968) considered the Cu in Cornish lodes to be derived by metasomatic removal from basic igneous rocks, in processes that need not directly have involved the granites. Edward (1976) concluded that Zn but not Cu could have been derived from primary concentrations in Devonian sediments of the province.

A concentration gradient in favour of granite against sediments indicates that the introduction of a particular element to the granites by assimilation is extremely unlikely. Elements showing this geochemical trend are likely to have been derived from the granites, or to have been introduced to them by some route other than assimilation, such as hydrothermal circulation. This trend can be observed in the concentration gradients of W, Ta, U, and Sn (Fig. 2). Tungsten and tantalum show a steady increasing trend from sediments through

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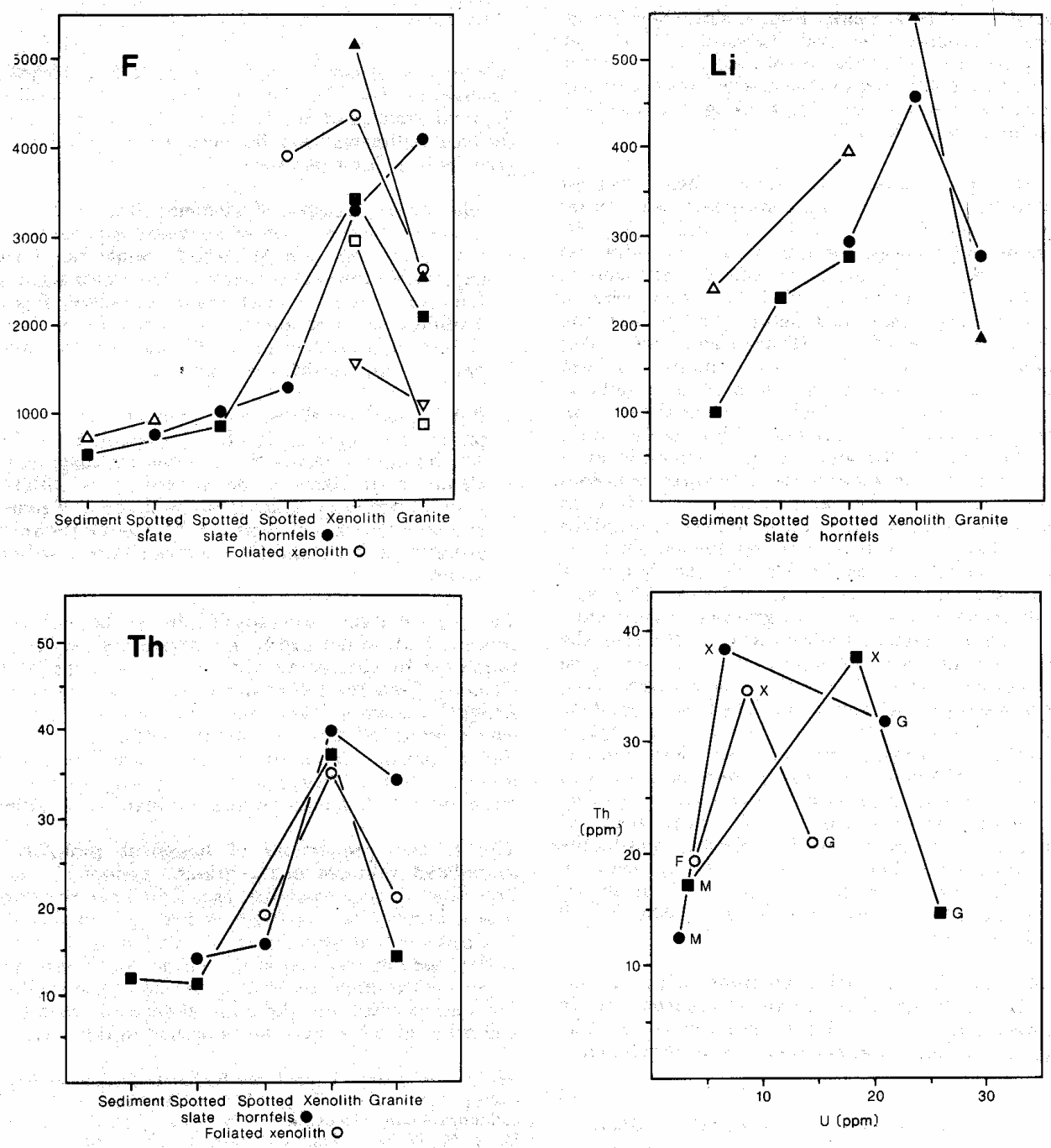

Figure 3. Geochemical gradients for fluorine, lithium and thorium (ppm; legend as for Figure 1), and variation in concentration of thorium vs. uranium (ppm; locality symbols as in Figure 1, and M = metasediment, F = foliated xenolith, X = xenolith, G = granite).

xenoliths to granites. Uranium shows a similar trend, with *very* low concentrations in sediments; however, uranium is well known to occur along with vanadium in some black shales, and since it is more mobile than vanadium it could be preferentially removed from black shales by circulating hydrothermal systems. Thus, a sedimentary source for uranium in south-west England cannot be precluded; although the present evidence suggests derivation from within the granite. Tin generally shows very low concentrations in sediments, the highest recorded Sn values in sediments outside the aureoles and away from the effects of mineralising fluids are about 5 ppm (although Edwards (1976) reported high Sn concentrations in some unmineralised Devonian sediments in Cornwall and concluded these could be a source for Sn). In most cases a steady increase in Sn from sediment through xenoliths to granite is observed, but at some localities, xenoliths contain more tin than their host granites, for example Chagford and Corndon Tor (both Dartmoor). Corndon Tor xenoliths are rather sandy in appearance, and are the only xenoliths of this type featuring in the study; they are geochemically anomalous in several ways, and it is quite possible that they were not derived from pelitic sediment. Tin concentrations are very variable

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within xenoliths from a single locality at Chagford, Corndon Tor and Luxulyan. Since no cassiterite has been observed in thin sections, the main host for tin in xenoliths is probably biotite (concentrations are, however, too low to detect using the electron microprobe) or possibly rutile.

Some elements consistently show their greatest concentrations in xenoliths by comparison with either granites or sediments, notably Li, Th and F (Fig. 3). These findings emphasise the need to analyse sediments as well as xenoliths and granites, in order to make sense of the results. The highest recorded Li values were in xenoliths (although published figures for the lithionite granites of the St. Austell area (Dangerfield *et al.*, 1980) are higher). The host mineral for lithium is almost certainly biotite, which is concentrated in xenoliths. Thorium also shows its highest concentrations in xenoliths, and can be attributed to inclusions within biotite; if thorium is the source of radiogenic haloes in biotite then it must be incorporated into these inclusions at a very early stage in the assimilation process, since, as already noted, the haloes first appear in foliated xenoliths similar in character to the contact sediments. There is, however, a marked contrast between the distributions of thorium and uranium, and the ratio of Th to U (Fig. 3) is higher in xenoliths than in either granites or sediments. Xenoliths are also preferentially enriched in fluorine; the xenoliths from all localities except Cape Cornwall gave higher fluorine values than either their surrounding granite or associated metasediment. The host mineral for F is again likely to be biotite, since very high F concentrations have been reported from biotites in the province (e.g. Al-Saleh *et al.*, 1977). Fluorine is likely to have been derived from the granite, but is evidently taken up by biotite recrystallising at an early stage of assimilation; this element is readily released from biotite by the slightest hydrothermal activity, and may then form fluorite (the present writer has observed secondary fluorite within partly chloritised biotites in the St. Austell granite (Lister, 1981).

Other elements concentrated in xenoliths include Ti, Zr, Rb and REE, although Ti and Zr are also concentrated in the contact aureoles, and Rb is also concentrated in granites, particularly in association with K-enrichment.

Pelitic Sediments	Metamorphic Aureole	Xenoliths	Granite
V	Ba	F	Sn
Cr	Sr	Li	W
Mg	Cu	Th	Ta
Mn	Zn	Rb	U
Zn	Ti	Ti	Rb
Ni	Zr	Zr	B
		REE	Be

Table 1. Trace element distribution in south-west England.

Conclusions

Assimilation of pelitic sediment cannot be regarded as a mechanism for the introduction to the south-west England granites of Sn, W, U, or Ta, but the nature of the local sediments has influenced the composition of the granites in at least two ways.

1) The direct transfer of elements from sediment to granite by assimilation of xenoliths is possible in the case of V, Ba and Sr, which, while not forming exploitable mineral deposits in their own right, are, however, found concentrated in some types of stanniferous tourmalinised veins within the granite (Lister, unpublished data). Gu and Zn may also be derived from outside the granites.

2) Biotite-rich xenoliths and contact metasediments provide a host for Li, F, Th, and possibly Rb and REE which might otherwise escape from the magma; these elements are likely to be derived from within the granite, but recrystallising biotite in the roof zone acts as a trap for these elements. Where xenoliths are then assimilated, Li, F and Th are thus retained within the granite.

The present data relate specifically to the assimilation process, and do not preclude sedimentary contributions to granite by circulating hydrothermal systems leaching elements from the sedimentary pile. The low grade of regional metamorphism, and the nature of sediments which remained hydrous, despite deformation, at the time of granite emplacement, could also have favoured mineralisation processes in the way suggested for the mineralised Caledonian granites by Plant *et al.* (1985).

The relative proportions of basement protolith and assimilated sediment in the granites cannot be assessed from the evidence available, especially since the original bulk compositions may not have been greatly dissimilar. It is possible that assimilation of upper crustal sediments of the type now exposed at the surface could represent a major contribution to the composition of the south-west England granites, but the unusual enrichment in Sn, W and other metals cannot be explained in this way.

Acknowledgements. I thank my former colleagues at Imperial College, particularly P. J. Watkins, for their guidance and collaboration on geochemical techniques. Thanks are also due to Dr. J. N. Walsh at King's College, London, and to Dr. S. J. Parry at Imperial College Reactor Centre for the use 0f facilities. This work was conducted during the tenure of a N.E.R.G. post-doctoral research fellowship, which is gratefully acknowledged.

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Read at the Annual Conference of the Ussher Society, January 1984

Field relationships within the South-west Jersey Granite Complex

A. M. BLAND

Bland, M. A. 1984. Field relationships within the South-west Jersey Granite Complex. *Proceedings of the Ussher Society, 6,* 54-59.

The South-west Jersey Granite Complex is intruded into Brioverian sedimentary rocks. Three main granite types are distinguished: (1) a coarse-grained granite; (2) a fine-grained microgranite; and (3) a porphyritic granite. These three lithologies form an annular outcrop pattern but field evidence, including the contact relationships, does not allow the relative ages to be determined. Using field evidence in conjunction with major element geochemical data, it can be shown that a simple comagmatic relationship between all three granites is unlikely. Isotopic data further indicate that the , coarse granite is not magmatically related to the other two types.

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Introduction

Jersey is the largest and southernmost of the Channel Islands and lies 30km off the coast of Normandy. Geologically the Channel Islands are more closely related to Brittany and Normandy than to the British Isles. They consist of a metamorphic basement of highly-foliated gneissose rocks of dioritic, granodioritic and granitic composition, which has been correlated with the Pentevrian basement of the French mainland, and comparatively unmetamorphosed sedimentary and volcanic rocks of Late Precambrian or early Cambrian age, which have been correlated with the Brioverian succession of the French mainland (Roach 1977). Throughout the Channel Islands, these earlier rocks are intruded by a variety of plutonic igneous complexes, some of which show a weak foliation. The region has been subjected to several major deformational episodes (Bishop *et al,* 1975).

The geology of Jersey has been examined by a number of workers in the past, the main articles relevant to this work being those of Henson (1950, 1956) and Adams (1967, 1976). Jersey consists, in part, of a series of sedimentary rocks and overlying andesitic and rhyolitic lavas. Emplaced into these are three plutonic igneous complexes Which now form the north-western, south-western and south-eastern corners of the island. In the north-west and south-east these plutonic complexes consist of rocks of granitic composition which are associated with earlier, more basic rocks of gabbroic and dioritic composition. In contrast,, the plutonic complex exposed in south-west Jersey consists entirely of rocks of granitic composition and there is no evidence of any associated more basic material.

This paper describes the field relationships of the granite complex of south-west Jersey and shows that these relationships present inconclusive evidence as to the relative ages of the constituent granites. It goes on to show that despite petrographic and geochemical similarities between different rock types, isotopic data indicate that the south-west Jersey granite may be more complex in nature and origin than has hitherto been appreciated.

N.B. Throughout this paper, map references are to the Universal Transverse Mercator grid on the 1:25000 O.S map of Jersey.

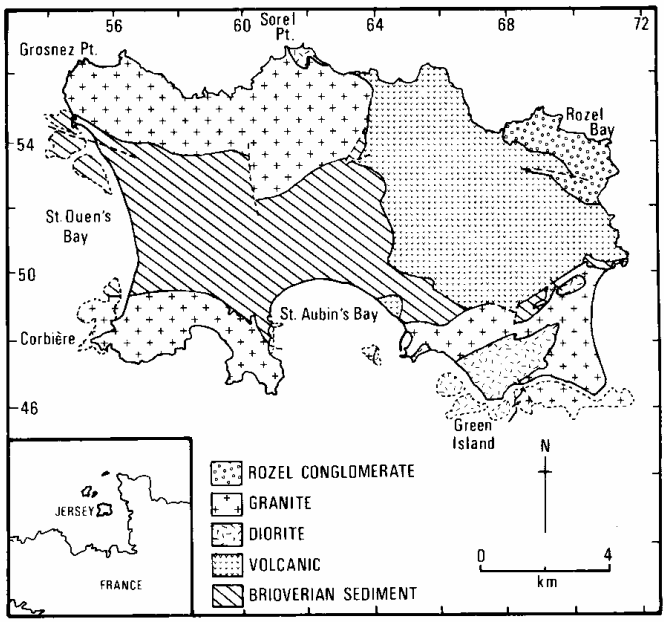

Figure 1. Simplified geological map of Jersey.

Field Relationships

The south-west granite extends from St. Ouen's Bay in the west to St. Aubin's Bay in the east (Fig. 1), covering an area of approximately 8km2. Exposure is excellent in coastal areas but poor inland where the topography is relatively flat and there are few road cuttings. The complex is intruded into the Jersey Shale Formation (Helm, 1983), part of the Brioverian succession. These sedimentary rocks include cleaved mudrocks, siltstones and greywacke sandstones. In general, the sedimentary rocks adjacent to the granite have been baked, some showing the development of cordierite. Sedimentary structures are preserved and can be seen for example in St. Ouen's Bay (GR560495). This low grade of contact metamorphism may be unrepresentative of the conditions

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at the time of intrusion since, at least at the eastern margin in St. Aubin's Bay, severe cataclasis of the granite suggests that the contact may be faulted. Only one outcrop occurs where an apparently untectonised contact may be seen but, unfortunately, this is a highly weathered inland exposure. Wherever the granite is seen close to the sediments, it is petrographically and geochemically indistinguishable from the same type of granite nearer the centre of the intrusion, although in the cataclased zone in St. Aubin's Bay partially assimilated xenoliths of sediment are present.

Four different granite types were recognised in the south-west complex by Henson (1950):

1. a marginal granite adjacent to the Brioverian sedimentary rocks

2. an early biotite-hornblende granite

3. a late microgranite

4. a porphyritic granite

The last of these, the porphyritic granite, was found only in association with the microgranite and was considered by Henson to be altered biotite-hornblende granite. Henson's division of the granite complex was used during the early stages of the present work but was found to be unsatisfactory in a number of respects. For example, the marginal granite is not as distinct a rock type as Henson suggests and, although the other three rock types can be recognised, both the microgranite and the porphyritic granite are inhomogeneous and can be further divided.

Three main granite types are distinguished here (Fig. 2): (1) a coarse-grained, reddish-weathering granite which forms most of the area of outcrop and is the only one of the three in contact with the country-rock; (2) a fine-grained, variably-textured, pink-weathering micro-granite which forms several isolated outcrops towards the centre of the complex; and (3) a brownish-weathering, porphyritic granite which outcrops in areas that lie between the coarse granite and the microgranite. Figure 2 shows that these three granite types have an annular outcrop pattern. The three granite types and their field relationships are described below.

Coarse Granite

Approximately 75 % of the exposed part of the south-west complex consists of the coarse granite which forms the outermost zone of the annular structure. This is an inequigranular, leucocratic rock showing variations in colour and texture. The coarse granite consists of 3-5% biotite, 0-3% hornblende, 30-45% perthitic alkali feldspar, 15-30% oligoclase, 25-35% quartz and accessory amounts of chlorite, zircon, epidote, apatite, ilmenite, allanite, fluorite, sphene and muscovite. The average grain size is approximately 4mm, although crystals greater than 15mm across are not uncommon. In any one hand specimen a complete gradation of grain size

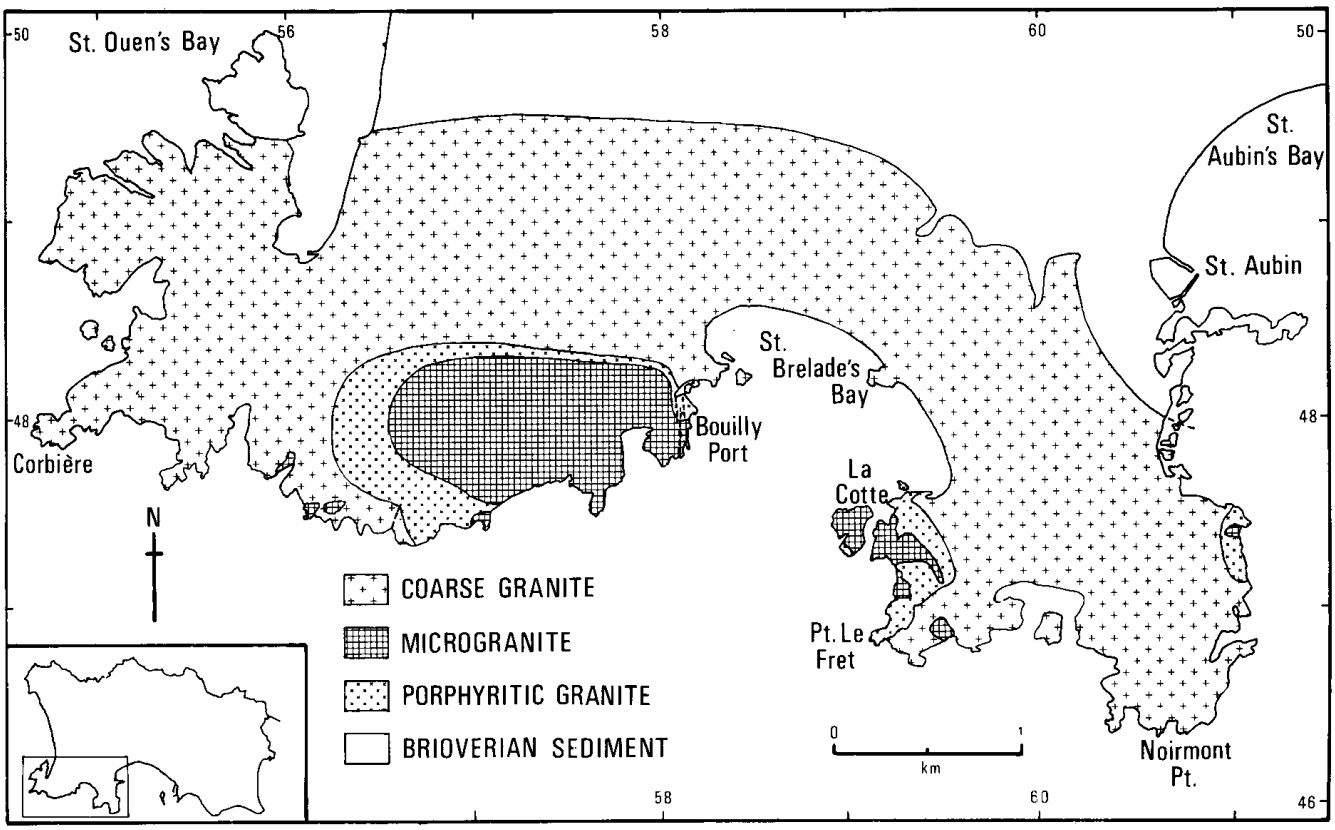

Figure 2. The South-west Jersey Granite Complex.

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may be seen, with quartz, alkali feldspar and oligoclase usually occurring as the largest crystals. It has frequently been suggested that large crystals may be developed metasomatically in granitic rocks (Stone and Austin 1961; Mehnert and Busch, 1981). In the case of the coarse granite of south-west Jersey, however, this seems an unlikely hypothesis, because a graphic texture is frequently developed in optical continuity with the large quartz crystals and graphic textures are generally considered to have a primary igneous origin (see, e.g. Cox, Bell and Pankhurst, 1979, p301). The implication is that the large quartz crystals are magmatic, and textural relationships (euhedral shape, general lack of inclusions), together with microprobe studies (A. M. Bland, unpublished data) suggest that this is also the case for the large alkali feldspar and plagioclase crystals. The relationship between the two feldspars is also of interest in this respect, since occasionally they occur together in a rapakivi texture, with white rims of oligoclase of varying thickness surrounding the larger pink perthitic alkali feldspar, for example at La Cotte (GR593476) and to the north of Noirmont (GR609466). Microprobe studies show that the rapakivi rims are different in composition to the perthite strings but are similar in composition to the outer portions of the zoned oligoclase crystals.

Microgranite

The second type of granite that is recognised in this work is the fine-grained, pink microgranite which forms several isolated outcrops towards the centre of the complex. It is well exposed at Bouilly Port and to the south of La Cotte (Fig. 2). It has a very low modal mafic content (less than 1%) and an average grain size of less than lmm. The microgranite is equigranular and consists of 35-40% perthitic alkali feldspar, 20-30% oligoclase, 35-45% quartz and minor amounts of biotite and ilmenite. Chlorite, zircon, apatite, fluorite and muscovite are accessory phases. Most crystals are anhedral and a graphic texture is commonly developed in the groundmass. Textural evidence suggests that the three felsic phases crystallised simultaneously. The microgranite shows a considerable variation in texture. Towards the centre of each area of outcrop it has an essentially non-porphyritic, equigranular texture, containing rare anhedral megacrysts of quartz up to 5mm across. The microgranite shows an increase in the percentage of subhedral and anhedral megacrysts of quartz, perthitic alkali feldspar and oligoclase towards the contact with the porphyritic granite. These megacrysts are larger than those seen in the non-porphyritic areas, and increase in amount to a maximum of 25%-30% of the rock in areas adjacent to the

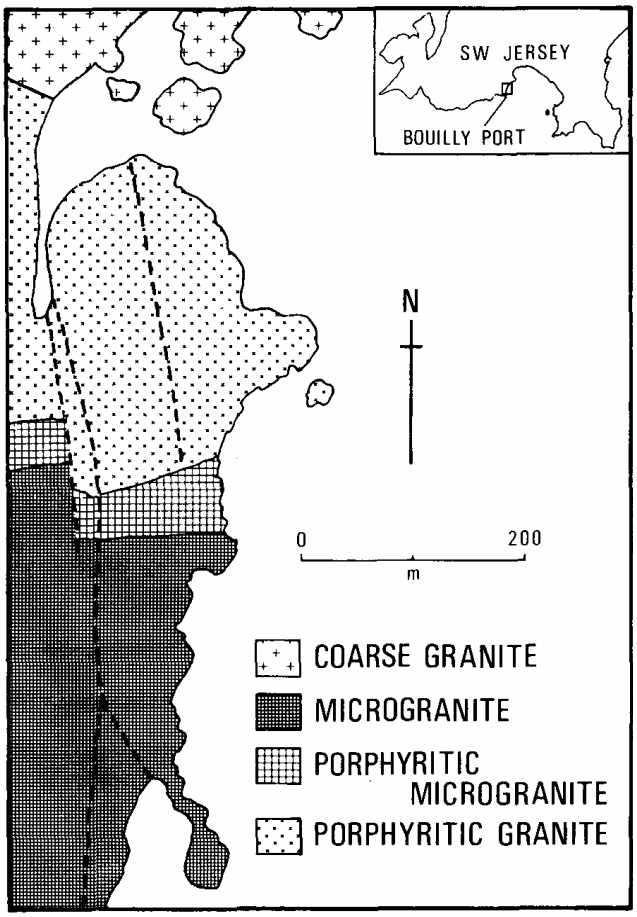

Figure 3. Geological map of the Bouilly Port area, S.W. Jersey

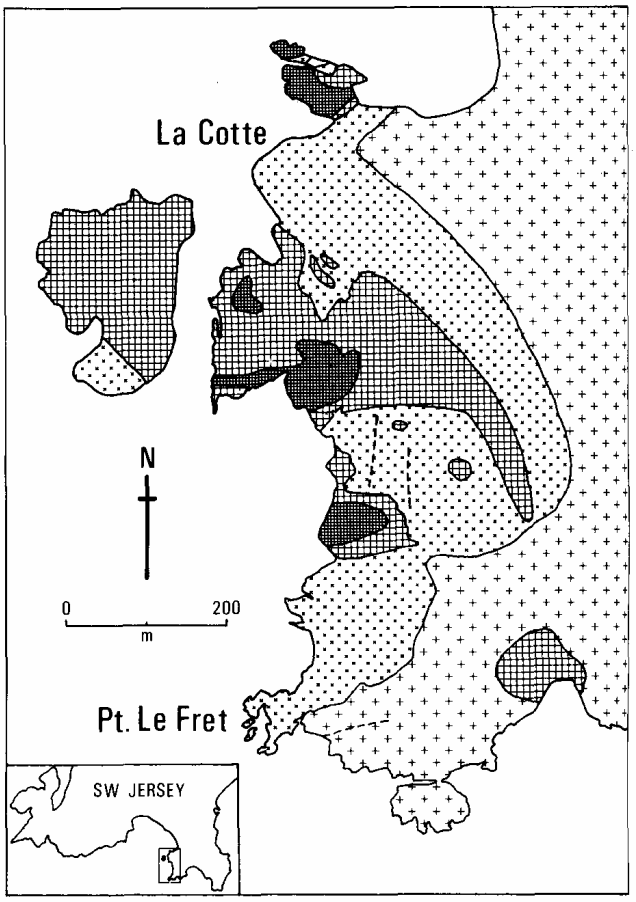

Figure 4. Geological map of the area to the south of La Cotte, S.W. Jersey. (Key as for Fig. 3).

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porphyritic granite contact. The rock in such areas is described as porphyritic microgranite and often may be mapped separately as a distinct variant of the microgranite. The microgranite has been divided up in this way in the coastal sections at Bouilly Port (Fig. 3) and to the south of La Cotte (Fig. 4).

Porphyritic Granite

The third main type of granite recognised in this work, the porphyritic granite, outcrops in areas that lie between the coarse granite and the microgranite, for example at Bouilly Port (Fig. 3) and in the area to the south of La Cotte (Fig. 2). It is red-brown in colour and has a well-developed porphyritic texture, commonly containing perthitic alkali feldspar, 20-35% oligoclase, 30-40% quartz, 0-2% ilmenite and, as accessory phases, chlorite, hornblende, allanite, epidote, zircon, sphene and apatite. A graphic texture similar to that seen in the coarse granite is developed, again suggesting that the megacrysts of quartz, at least, are of primary igneous origin. As in the-case of the coarse granite, the euhedral shape and textural relationships shown by many of the feldspar megacrysts indicate that they too may be of primary rather than metasomatic origin. A rapakivi texture is developed in some areas. The porphyritic granite exhibits a regular variation in texture which is related to the proximity of the other two granites. In areas adjacent to the coarse granite, for example at La Cotte, the porphyritic granite consists of 90% megacrysts and 10% groundmass. With increasing distance from the coarse granite the number of megacrysts decreases until adjacent to the microgranite the porphyritic granite contains 70% groundmass and 30% megacrysts.

Contact Relationships

The contacts between the porphyritic granite and the other two granite types do not exhibit simple intrusive relationships.

The area of contact between the porphyritic granite and the coarse granite is well exposed at La Cotte (Fig. 4, GR593476), where a gradation from porphyritic to non-porphyritic granite can be seen over a distance of 0.5m. In this 0.5m interval the granitic rock shows characteristics intermediate between those shown by the porphyritic granite and the coarse granite. In particular, the lithology shows a bimodal distribution of grain size, which does not normally occur in the coarse granite, but the percentage of groundmass, about 10%, is less than that normally shown by the porphyritic granite. In each such area of contact, the boundary between the two granites appears to be relatively planar and no lobing of one into the other can be seen.

The porphyritic microgranite/porphyritic granite contact is more distinct, and is well exposed at Bouilly port and to the south of La Cotte. It can be placed to within approximately 10mm and occurs where there is a distinct colour 'change from pink porphyritic microgranite to brown porphyritic granite, and a corresponding increase in the percentage of mafic minerals, for example to the south of La Cotte (GR593471). In the coastal section at Bouilly Port (Fig. 3) a number of changes can be seen in a traverse northwards from the non-porphyritic microgranite at the southern end of the section through porphyritic microgranite and into porphyritic granite:

(1) A gradual increase in the number and size of felsic megacrysts, and the appearance, in the porphyritic granite, of biotite megacrysts.

(2) An increase in the number and percentage of mafic phases in the groundmass, from the microgranite which contains biotite only, to the porphyritic granite which contains biotite, hornblende and ilmenite.

(3) A decrease in the percentage of groundmass from greater than 99% in the microgranite to less than 55% in the porphyritic granite.

Although contacts can be seen between porphyritic microgranite and porphyritic granite, it is not normally clear in which order the two were emplaced; contacts are usually more or less planar and neither lithology shows a chilled margin. However, to the south of La Cotte (GR593471), two pieces of evidence which demonstrate the relative ages of the two lithologies can be found. Firstly, a xenolith of porphyritic granite measuring some 200mm across occurs within the porphyritic microgranite approximately 250mm from the main granite contact. Secondly, a lobe of porphyritic microgranite extends into the porphyritic granite for lm, indicating that the latter is the older.

The age relationship between the microgranite and the coarse granite is indeterminable in the field, since these two are normally separated by porphyritic granite. Thus, from field evidence alone, it is not possible to place all three main granite types in chronological order.

There is, however, one cross-cutting relationship that can be demonstrated within the complex. In several areas, for example at Corbiére, planar sheets of a non-porphyritic microgranite, up to lm thick, cut across the coarse granite. The upper margin of each sheet normally has a relatively sharp contact with the overlying coarse granite. In contrast, at the base of a sheet, the microgranite grades downwards into a porphyritic granite which, it should be emphasized, is different in character to the main porphyritic granite described earlier. Approximately lm below the base of the microgranite sheet the porphyritic granite grades downwards into normal coarse granite over a distance of a few tens of millimetres. This pattern is typical of most microgranite sheets, although a few consist only of microgranite, and have sharp upper and lower contacts with the coarse granite. There are no such sheets within the porphyritic granite. Thus, although these sheets of microgranite and associated porphyritic granite can be shown to post-date the coarse granite, their relationship to the large areas of microgranite and porphyritic granite is uncertain.

In several areas, for example at Noirmont Point and at Corbiére, the coarse granite contains irregular patches of microgranite a few metres across which are sometimes associated with a porphyritic granite. Where topography is fiat it is impossible to ascertain whether these are intrusive sheet-like structures within the coarse granite or whether

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they are included as xenoliths; given the presence of microgranite sheets at Corbiére, the former interpretation seems more likely.

General Features

There are a number of features that are common to all three granite types.

(1) Cataclastic bands can be seen in many areas; these trend dominantly WSW-ENE with a minor NW-SE component. The bands vary in thickness from less than 0.5mm to tens of millimetres; cataclased zones. containing a number of poorly defined bands, can measure more than lm across. Individual cataclastic bands consist largely of fine-grained, grey-black, fragmented granitic rock in which relatively resistant quartz crystals have been preserved as porphyroclasts. The intensity of cataclasis varies considerably but is generally least in the microgranite and greatest in tile coarse granite.

(2) Aplitic and microgranitic dykes and veins of various thicknesses are common and are sometimes accompanied by pegmatitic quartz and alkali feldspar. These dykes trend dominantly E-W and, more rarely, N-S; some aplitic dykes are demonstrably older than some of the basic dykes (see (4), below) but it is not possible to ascertain whether or not the aplite dykes are all of one generation.

(3) Joints are particularly well-developed in the microgranite; contoured stereograms show three dominant orientations: sub-horizontal; N-S sub-vertical; and WNW-ESE sub-vertical.

(4) Basic dykes, mostly doleritic, more rarely lamprophyric, intrude the granites and adjacent sediments. The dykes vary in thickness from tens of millimetres to several metres and trend dominantly WSW-ENE and NNE-SSW. Cross-cutting relationships show there to have been at least two episodes of dyke emplacement.

(5) Minor faults of various orientations occur over much of south-west Jersey, usually striking either N-S or E-W. Slickensides are rare, and the sense of movement can normally be determined only where dykes or veins are displaced.

Geochemistry

For the South-west Jersey Granite Complex, most of the major oxides (Al_2O_5 TiO_2. Fe_2O_5, MgO, MnO, CaO and P_2O_5) show regular trends, decreasing in amount with increasing SiO_2 Na_2O and K_2O produce more scattered plots. Whilst the origin of these trends is not debated here and does not affect the interpretations made, it is clear that, in all of these variation diagrams, the samples fall into groups which correspond to the different rock types recognised in the field. A plot of Al_2O_5 vs. SiO_2 shows these groupings (Fig. 5). A total of 77 samples is plotted. The non-porphyritic microgranites (closed circles) have the highest SiO_2 values and are geochemically indistinguishable from the microgranite sheets (open circles). The porphyritic microgranites (closed diamonds) are slightly less SiO_2 rich but plot adjacent to the

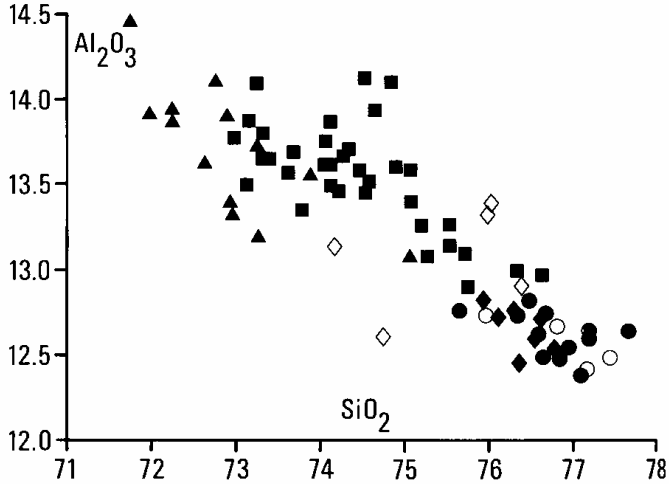

Figure 5. Al_2O_3 vs SiO_2. Symbols: triangles—Porphyritic Granite; squares—Coarse Granite; closed circles—Nonporphyritic Microgranite; closed diamonds — Porphyritic Microgranite; open circles—Microgranite Sheets; open diamonds— Porphyritic Granite associated with Microgranite Sheets.

microgranites. The porphyritic granites (triangles) have the lowest SiO_2 values and plot as a group which is distinct from the porphyritic granite associated with the microgranite sheets (open diamonds); the latter plot as a group adjacent to the porphyritic microgranites. The coarse granites (squares) have a relatively wide range of SiO_2 values and plot between the two types of porphyritic granite. A number of conclusions can be drawn from variation diagrams such as this:

(1) The main porphyritic granite plots in such a position that it is unlikely to have formed as a result of mixing between coarse granite and microgranite, which both plot to the SiO_2 -rich side of the porphyritic granite.

(2) The main microgranites and microgranite sheets have a similar major element chemistry but it does not necessarily follow that they are related to the same period of intrusion.

(3) The porphyritic granite associated with the microgranite sheets is not geochemically the same rock type as that which forms the main areas of porphyritic granite but plots in such a position that it could be the result of a mixing between coarse granite and intruding microgranite sheets.

Groups of rocks which lie on a simple linear or curvilinear trend on variation diagrams are often interpreted as belonging to a single, comagmatic suite, and are frequently assumed to be related to each other by a process such as fractional crystallisation. In general, it might be expected that the oldest rocks would be those that are least evolved and have the lowest SiO2 values. If the granites of south-west Jersey, which show a single trend, are related by fractional crystallisation, one might suppose that the groups would be emplaced in an order that is related to the annular structure of the complex and that the geochemical variation would be consistent with this emplacement order. Given the outcrop pattern, this would mean that either the coarse granite or the

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microgranite should be the least evolved and have the lowest SiO_2 values The plot of Al_2O_3 vs SiO_2 shows that this is not the case. It is clear from this plot that the porphyritic granite, which is the middle of these three granites in the annular structure, is the least evolved, the microgranite the most, and the coarse granite plots between the two. It is suggested, therefore, that a simple comagmatic relationship between all three granites is unlikely.

Isotopic Data

It seems from both the field and geochemical evidence that the South-west Jersey Granite Complex did not form from a single evolving magma, despite the fact that the simple annular structure of the complex might suggest this to be the case. New Rb/Sr isotope data lend further support to the view that the three granites are not cogenetic. The age of the south-west Jersey granite has previously been quoted as 565 ± 15 Ma (Adams, 1967, using a decay constant of 1.39×10^{-11} year $^{-1}$ for ^{87}Rb), based on a Rb-Sr isochron from a suite of 5 samples (4 whole-rock and 1 biotite) which was almost certainly a mixture of what are here described as coarse granite and porphyritic granite. Recently determined separate isochrons for each of the three main granite types, using a decay constant of 1.42×10^{-11} year $^{-1}$ for ^{87}Rb, (A.M. Bland, unpublished data--to be presented in detail elsewhere) show that this single age probably has little significance. Although the microgranite (527 ± 7 Ma) is close in age to the porphyritic granite (550 ± 12 Ma), the coarse granite (483± 13 Ma) is considerably younger than either of these two. At the limits of error the microgranite and porphyritic granite give ages within 4Ma of each other. Taking into account the geochemical relationships, the possibility that the microgranite is a magmatic derivative of the porphyritic granite cannot be entirely discounted. In contrast; the isotopic data argue strongly that the coarse granite is not related to the other two types.

Conclusions

(1) Given the relative ages of the porphyritic granite and the microgranite, and considering currently available geochemical data, it remains a possibility that the microgranite could be a magmatic derivative of the porphyritic granite.

(2) Compared to the porphyritic granite and the microgranite, the coarse granite is: a) structurally the outermost; b) geochemically intermediate; c) much the youngest. Taken together, these observations strongly suggest that the coarse granite is not related to the other two types, despite petrographic similarities between the coarse granite and the porphyritic granite and despite the gradational contact that exists between these two.

(3) A suite of microgranite sheets post-dates the coarse granite. It follows from the isotopic data that these sheets cannot be related to the main microgranite.

(4) The minor amounts of porphyritic granite associated with the microgranitic sheets are similarly unrelated to the main porphyritic granite, but may be produced by some form of "mixing" between coarse granite and late microgranite sheets.

(5) All available evidence, including unpublished microprobe data, supports the view that the megacrysts are of primary igneous origin.

(6) The thesis of Henson (1956), which suggests that the porphyritic granite was formed as a result of metasomatic alteration of an outer, supposedly early, biotite-hornblende granite (the coarse granite) during the emplacement of the central microgranite, is untenable in the light of all the available evidence.

Acknowledgements. This work forms part of a Ph.D. study funded by an Oxford Polytechnic Research Assistantship, which is gratefully acknowledged. Thanks are due to Dr. M. Brown, Dr. G. M. Power and Mr. C. G. Topley, the supervisors of this project, who have given helpful advice throughout the study and, particularly Mr. C. G. Topley, during the preparation of this manuscript. I am indebted to the Department of Geology and Mineralogy at Oxford University and the Department of Geology at Portsmouth Polytechnic for the use of their facilities.

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Read at the Annual Conference of the Ussher Society, January 1984

Hercynian high-angle fault zones between Dartmoor and Bodmin Moor

P.J. TURNER

Turner. P. J. 1984. Hercynian high-angle fault zones between Dartmoor and Bodmin Moor *Proceedings of the Ussher Society 6,* 60-67.

The nature of the dominantly NNW-SSE trending high-angle fault zones between Dartmoor and Bodmin Moor is described. These structures are demonstrably of Hercynian age and possibly older, with little or no Tertiary displacement. The fault zones are here interpreted as boundary zones within and between northward translated nappes which were thinning by gravitational spreading. The zones show little evidence of wrench movements.

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Introduction

The occurrence of NNW-SSE trending fault zones has long been recognized in south-west England and their significance in the structure of the region has been reviewed by Dearman (1963). He concluded that they were essentially dextral wrench faults of Tertiary age. Other authors (Blyth 1957, Lane 1969, Freshney *et al.* 1972, Freshney 1977, Hamblin 1972) have suggested that at least some of the fractures may have been initiated before the Tertiary, but no detailed descriptions of the fault zones has previously been undertaken. The view that the fault zones were extant by the Hercynian is strongly supported by work recently completed between Dartmoor and Bodmin Moor during the revision of B.G.S. 1:50,000 Sheet 337 (Tavistock). This region is now recognised as a nappe and thrust terrain (Isaac *et al.* 1982, 1983), in which autochthonous and parautochthonous successions have been overridden by nappes derived from the south. Over much of the area the Kate Brook Unit occurs at the lowest structural level and may be autochthonous; this is overthrust, in upward sequence, by the Greystone, Petherwin, Tredorn and Blackdown Nappes (Fig. 2b and Table 1). The relationship of the Boscastle Nappe to the latter is not well constrained and remains to be elucidated. In general, from east to west progressively higher tectonic units are encountered across NNW-SSE trending fault zones. Because published B.G.S. maps have failed to distinguish formations adequately, these fault zones have hitherto largely passed undetected; one skirts the eastern margin of Bodmin Moor, another aligns broadly with the River Tamar and a third extends between Stowford and Lamerton (Fig. 1). It now appears that these faults played a critical role in the evolution of the nappe structures; this role is explored here.

The fault zones are generally poorly exposed. However, during the course of this investigation extensive trenching took place at a number of localities along the Tamar Fault Zone: 2km of trenches were opened around Eastcott Farm (SX3879) and this was supplemented by a planned programme of pit excavation. Additionally, over 100 boreholes in the Wrixhill (SX3779) and Greystone (SX3680) areas were generously made available by English China Clays (Quarries Division), together with access to their working quarry. Trial pits and boreholes

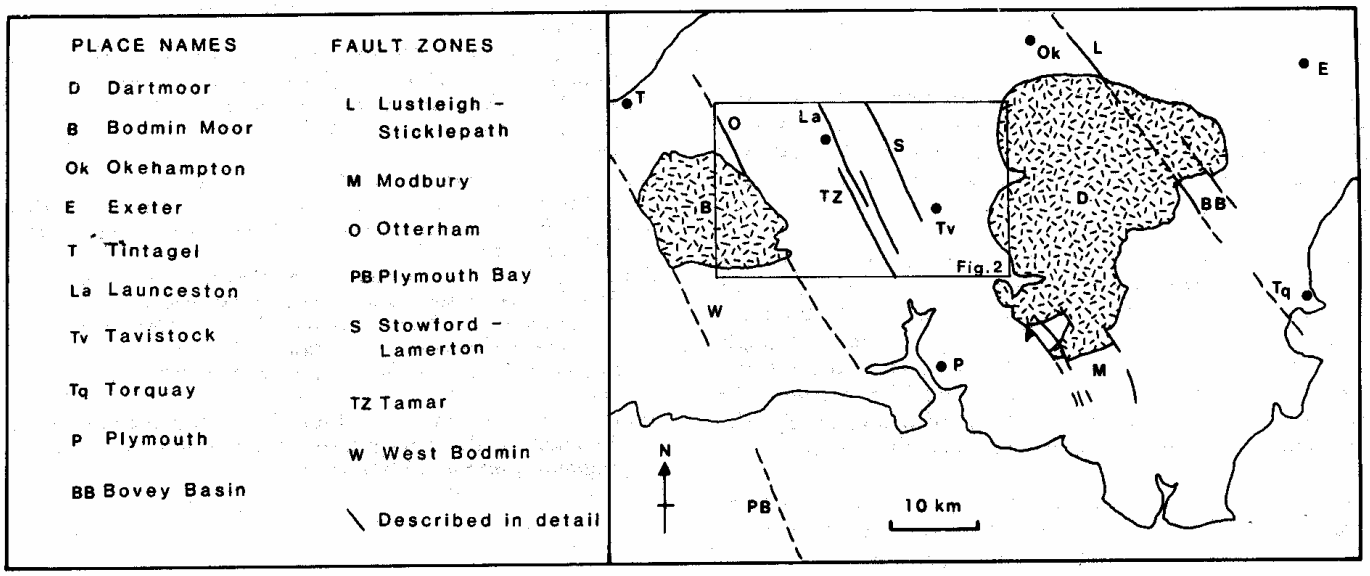

Figure 1. Location map of fault zones discussed in this paper

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D3 Klippen	Turner (1982b, in preparation)
Boscastle Nappe	Selwood *et al.* (in press)
Blackdown Nappe	Isaac *et al.* (1983) (Includes Bealsmill Formation, Turner 1982b, (in preparation) and St. Mellion Outlier, Whiteley 1981)
Tredorn Nappe	Isaac *et al.* (1982, 1983)
Trekelland Thrust	Stewart (1981a), Isaac *et al.* (1982)
Petherwin Nappe	Isaac *et al.* (1982, 1983)
Greystone Thrust	Isaac *et al.* (1982, 1983), Turner 1982a, b)
Greystone Nappe	Isaac *et al.* (1982, 1983), Turner (1982a, b) (Includes Lydford Unit (Isaac 1981, Turner 1981), Chillaton accounts (Turner 1981, Selwood *et al.* 1982) and Heathfield Nappe (Isaac *et al.* 1982, 1983))
Main Thrust	Isaac (1981), Turner (1981), Isaac *et al.* (1982)
Parautochthon	Turner (1982b, in prep.) (Includes Liddaton Unit (Isaac 1981, Turner 1981)
Whitelady Thrust	Isaac (1980, 1981)
Kate Brook Unit	Isaac *et al* (1982, 1983)

Table 1. Tectonic Units recognized between Dartmoor and Bodmin Moor

from the A30 have also provided subsurface data at the northern end of the fault zone. It has thus been possible to determine the character of the geology of this fault zone in great detail. An interpretation has been developed which can be satisfactorily applied to the adjoining fault zones.

Tamar Fault Zone

This zone is described between the northern margin of the Gunnislake Granite (SX4272) and the Tertiary basin at Dutson (SX3486) (Fig. 2). It is generally up to 2km wide, but may locally increase to 3km. Although certain sections are marked by deeply dissected topographic lineaments, the zone is principally recognised by the juxtaposition of contrasting tectonostratigraphic units (Fig. 2b). Thus west of the fault zone, and at its northern end, the Petherwin Nappe with overriding Crackington Formation is exposed, whilst to the east the allochthon (Greystone Nappe) terminates 3kin further south, and the Petherwin Nappe is not recorded to the east of the fault zone. Further south, in the Greystone area (Turner 1982a) and the Wrixhill area (Figs. 2 and 3) the Petherwin Nappe is juxtaposed with the Greystone Nappe. Clearly there is no simple relationship of downthrow to the west, although higher tectonic levels are preserved west of the zone. Across the eastern boundary fault of the zone, at Eastcott Farm, higher parts of the Greystone Nappe are preserved to the west of the fault than to the east. Immediately south of Wrixhill (Figs 2 and 3) the Bealsmill Formation (Blackdown Nappe) overrode the Greystone and Petherwin Nappes and filled a depression in the underlying thrust sheets. The flysch is presently juxtaposed with the Greystone Nappe along the western boundary of the fault zone. The changes in tectonostratigraphy generally take place across a single NNW-SSE trending fault within the zone. However, considerable complexity of structure is reflected in the interaction of faults within the zone. The principal faults which constitute the zone are described below.

NNW-SSE faults

The character of these faults, the most important elements in the fault zone, is best displayed about Greystone Quarry (Turner 1982 a, b). Some are steeply dipping normal faults, but others appear as parallel structures dipping steeply westwards, which become listric at depth. It has been possible to show (Turner 1982 a, b) that they form lateral ramps and undulatory thrust surfaces of D1 and D2 ages. Slickensides and small scale folds indicate that displacement at these interfaces was towards the north and they are interpreted as side wall ramps, developed during the emplacement of the nappe sequence. These side wall ramps may only persist for a few hundred metres, so that surface outcrops show discontinuous fault traces together with an apparent rotation of the fault boundary perpendicular to the main zone boundaries. This may lead to spurious assessments of displacements across NNW-SSE trending structures, particularly in poorly exposed ground with little topographic variation. Shear zones developed below the Petherwin Nappe were deflected into near vertical zones where side wall ramps were exposed in the quarry. Detailed cross sections, a thrust surface contour map, descriptions and photographs of these structures are given in Turner (1982 a, b).

The complex nature of the NNW-SSE trending faults was also revealed by extensive excavations around Eastcott Farm. These indicate that fractures resolved as single faults at 1:10,000 scale, are composed of anastomozing networks in which normal and oblique movements have produced local steepening of fabrics, brecciation and joint development, reflecting highly variable strain. Brecciation is generally associated with the redistribution of iron and manganese in vein networks which are truncated by the faults.

E- W and ENE- WSW faults

These elements, both normal and reverse faults, form local accommodation structures to the principal movements on the closely spaced NNW-SSE faults; as such they have diverse origins and ages. In the Bradstone area (SX3881) they originate from the interaction of listric reverse faults with the NNW-SSE trending faults, in the toe region of the Greystone Nappe (Turner, in prep.). Several formations became intercalated as the nappe thinned and terminated in this area and high-angle normal faults are intimately associated with the listric structures (Turner, in prep.). Gently inclined thrusts show a variety of orientations passing through the fault zone depending on the local topography. Of particular interest is the role of some of the thrusts, for example the thrust in the Tamar Woodlands area (SX3878) of the fault zone (Fig. 4 T). Fault F-F' does not displace the thrust, though structural disparities are revealed on either side of it. This fault was operating at the same time as the thrust; successions on either side appear to have moved northwards to different degrees relative to one another along the basal thrust· The thrust itself did not migrate, only the successions above the thrust surface, so that the normal limb of a north facing recumbent fold is preserved west of the fault and the lower limb and hinge region of

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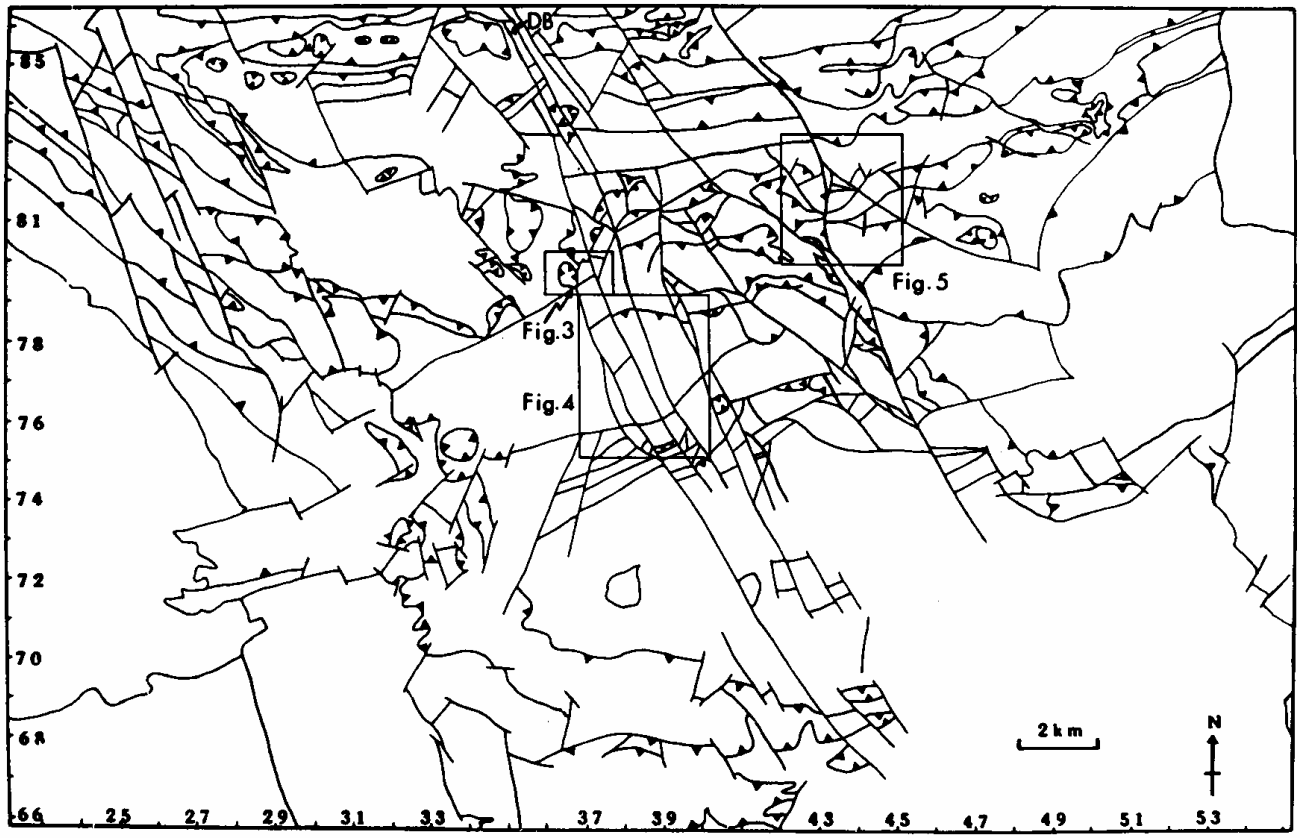

Figure 2a. Map showing principal geological boundaries between Dartmoor and Bodmin Moor

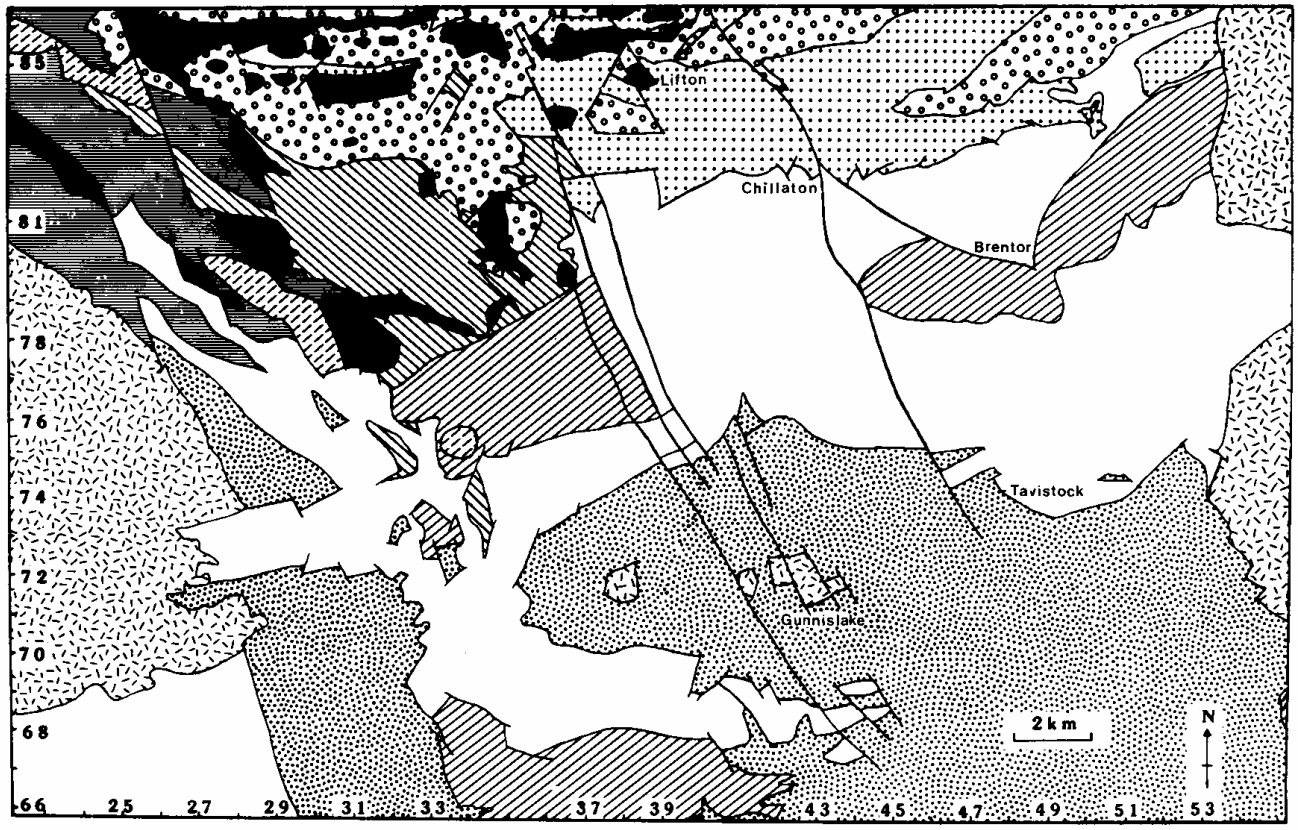

Figure 2b. Map showing principal tectonic units between Dartmoor and Bodmin Moor

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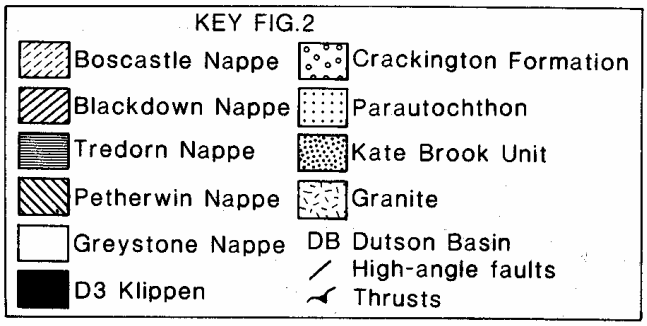
the same structure are preserved to the east. The lack of offset of the thrust precludes the possibility of any Tertiary displacement on this segment of the fault zone. Comparable structures are developed in the Lifton area at the northern end of the fault zone. A N-S section oblique to the zone (Turner, in prep.) illustrates the juxtaposition of formations across the faults within the zone. The D3 (Firebeacon) klippen are variably truncated by the faults (Fig. 2), which may appear to indicate late D3 or younger movements. However, the thickness of the klippen and the underlying thrust sheets is highly variable across the faults and there is no evidence of simple vertical displacement. The continuity of shallow dipping faults across these NNW-SSE structures precludes lateral displacement. The complex juxtaposition of the formations is best explained by invoking the interaction of these faults with the northward displaced thrust sheets of the allochthon and parautochthon during D1 to D3. The faults acted as complex boundary zones within and between the thrust slices and in some cases were not listric side wall ramps, but persist at depth (see also Figs. 3c and 4).

Mineralization

The Tamar Fault Zone is variably mineralized along its length and not solely where it crosses the main mineralized belt associated with the Gunnislake Granite. The cross course exposed in the Devon Great United Mine exploits the eastern boundary fault of the zone. Kaolinisation and fault gouges are variably developed along the length of the NNW-SSE trending faults in this area too (Bull 1982).

Early mobility of trace elements from pelites into bounding faults of the zone also appears to account for localized anomalies (particularly molybdenum) which caused severe stock problems at Brimble Brook Farm (SX4174). An investigation into this matter showed that a high concentration of toxic trace elements was associated with the mobility of silica during deformation. The highest molybdenum values occur where the slaty cleavage in the pelites is intensely crenulated and quartz veins, with associated mineral concentrations, invade the lithologies in complex networks.

Mineralization of the fault zone is associated with the redistribution of elements from the sediments and intrusives during the Hercynian deformation. The age and origin of the mineral assemblages is dependent upon the local stratigraphy and strain conditions in the evolving nappe sequence. Examples have been documented from excavations in the Greystone, Eastcott and Chillaton areas (Turner 1982 b).



Tertiary movements

Within the northern part of the Tamar Fault Zone, at Dutson (SX3486) (Fig. 2), a small Tertiary basin of fluvial origin has been identified (Freshney *et al.* 1982). The restricted nature of the deposit, and its probable sedimentary contact at its southern margin on the Upper Carboniferous (Freshney *et al.* 1982), militate against significant Tertiary fault movement. The possibility of lateral displacement on the NNW-SSE trending boundary faults having led to the development of a Tertiary "pull-apart" structure, is precluded by the lack of offset to Hercynian (D3) Klippen which straddle the faults in this area (Turner, in prep.). The evidence from the Palaeozoic strata suggest that little or no post Hercynian displacement has occurred across this fault zone further south.

The fault bounded margins to the Gunnislake Granite (Fig. 2b) provide no evidence of younger movements. The granite bodies in this area are of different types and Bull (1982) has also suggested that their emplacement was controlled by the faults during the Hercynian; their thermal aureoles being preserved across the faults.

Stowford-Lamerton Fault Zone

This zone, which has been traced from Stowford (SX4086) in the north to beyond Lamerton (SX4577) in the south, can be resolved, over much of its length, into a single fault (Fig. 2). It thus appears a less significant structure than the Tamar Fault Zone, nevertheless it shows comparable features. Basically two NNW-SSE segments are represented; one principally developed in the parautochthon in the north, offset from the other in the south by the N-S trending Hogstor Thrust (Figs. 2 and 5). The northern limit of last approximates to the E-W trending northern limit to the allochthon (Greystone Nappe) which crosses the fault zone vertically, without significant displacement (Figs. 2 and 5). Significant Tertiary wrench movement is thus precluded. Detailed investigations about the Hogstor Thrust have revealed contrasting formations to the east and west and a structure which preserves higher tectonic levels to the west. West of this thrust the successions are normal, but to the east they are mainly inverted. Between the Hogstor Thrust and the Chillaton-Brentor Splay, north closing folds in an essentially inverted sequence are recorded (Fig. 5). As in the comparable structures preserved in the Tamar Fault Zone, the Hogstor Thrust is interpreted as a side wall ramp, developed as an important thrust within the Greystone Nappe.

Interpretation

Both the Tamar and Stowford-Lamerton Fault Zones are directly associated with the interaction of displaced thrust sheets in the allochthon and parautochthon (Fig. 2b). Some of the faults in these zones are steep west or east

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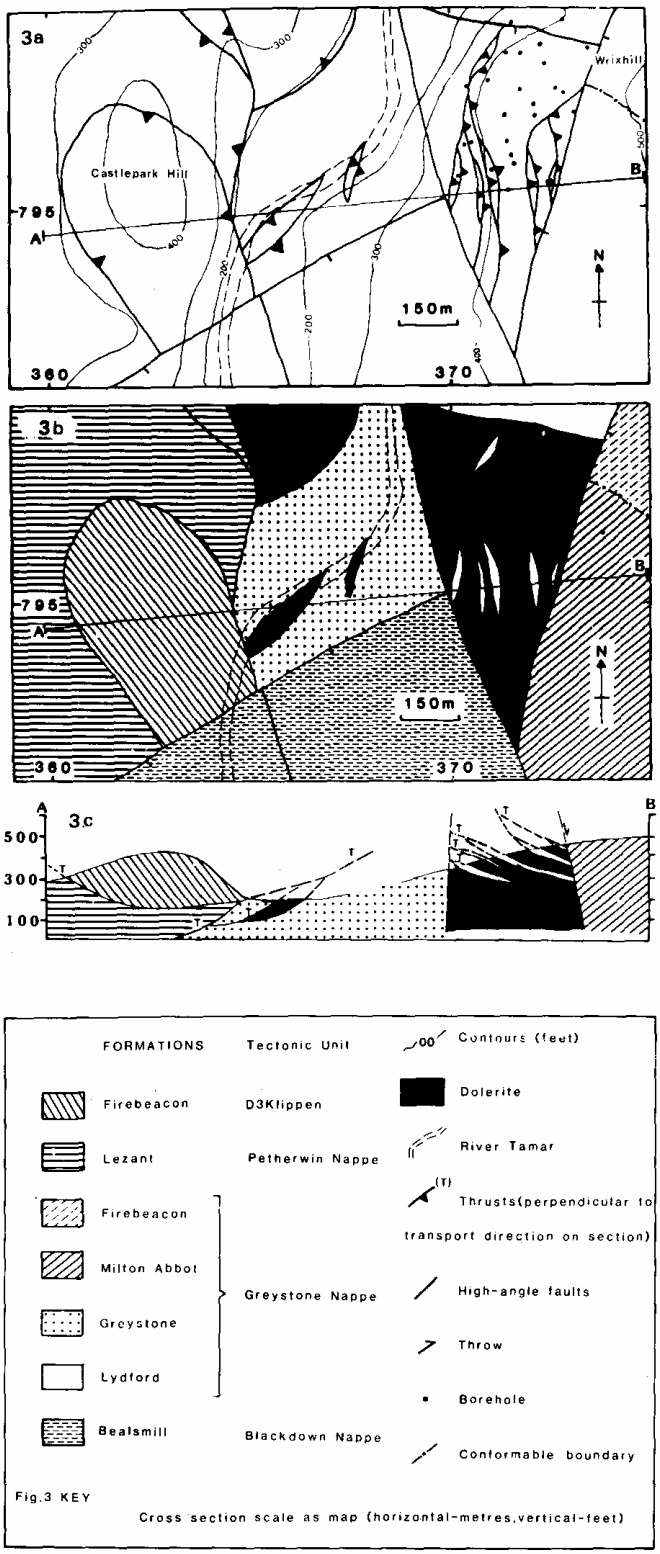

Figure 3a. Map of the Wrixhill area showing geological boundaries

Figure 3b. Map showing formations recognized in the Wrixhill area

Figure 3c. E-W cross section of the Wrixhill area

dipping thrusts along which oblique movement has occurred. Generally these form side wall ramps, of variable persistence, with movement northwards perpendicular to the dip of the surface. Folds are variably developed on either side of these structures and striking variations in thrust geometry across the faults are also seen. No simple components of displacement can be resolved across the faults. It appears that the development of high-angle faults, in the movement direction, permitted differential displacements of the thrust sheets. Such interactions explain the complex strain variations and contrasting structures observed between the faults.

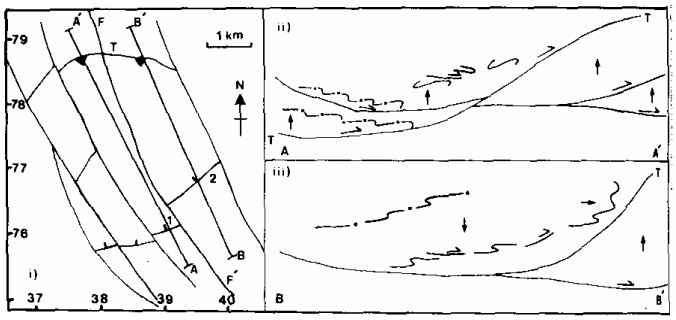

Figure 4. Schematic cross sections through the Tamar Fault Zone south of Wrixhill

This interpretation is consistent with that offered by Dubey (1980) to account for contrasts in fold style across high-angle faults (in the absence of lateral displacements) within the Bude Formation, immediately north of the area discussed in this paper. He described a modelling experiment in which the relationship between disparate structures across such faults was explained by their cognate development. Folding and faulting are intimately associated and stress may be transmitted through either, affecting the local development of the other type of structure. Once faults begin to propagate, the deforming cover (by analogy with the model) becomes divided into several regions in which the development of structures is different and independent of its surroundings. Some folds may have counterparts across the faults, others cannot be traced across the structures.

Discussion

Dearman's (1963) view of the wrench fault system of south-west England has proved particularly influential. Although a good many of his faults were inferred, the overall distribution of fault zones is broadly consistent with present day knowledge. What has to be questioned, in the light of this paper, is the assertion that they all represent dextral wrench faults of Tertiary age.

Evidence for Tertiary wrench movement is largely derived from the Lustleigh - Sticklepath Fault (Fig. 1); in particular, its displacement of the Dartmoor Granite and its well documented association with the Tertiary clays of the Bovey Basin (Edwards 1970). However this does not deny earlier Hercynian movements, indeed contrasts in tectonostratigraphy either side of the fault zone have been

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reported by Waters (1970, Fig. 2). Furthermore, Selwood and Thomas (1984) indicate that allochthonous units in the Belstone area are limited eastwards against autochthon by the Sticklepath Fault.

Stewart (1981 b) has described features of the Otterham Fault System (Fig. 1) (Selwood 1971), part of the Plymouth- Cambeak Fault Zone (Dearman 1963), NE of Bodmin Moor. He noted the vital role played by this zone in separating different tectonostratigraphic levels and suggested that granite emplacement was locally controlled by fractures in this zone. Tertiary movements are not recognized.

The fault along the western margin of Bodmin Moor also appears to have affected the emplacement of the granite at high levels (Bristow, personal communication). Again, there is no positive evidence of significant Tertiary movement. Lane (1969) has suggested that oblique slip was operative, at least at the south coast, where he described the interaction of folds with strike-slip movements on the Porthnadler Fault. Dearman (1963) extended this fault, along the western margin of Bodmin Moor, to the north coast. This fault appears to be a continuation of the Plymouth Bay Fault (Day and Edwards 1983) further to the SE (Fig. 1). These authors explained the lack of seismic reflectors to the east of the fault as a result of Tertiary displacement along the fault and cite Dearman's proposed 54km of dextral displacement. This figure was derived using stratigraphic information on maps available at the timer The presence of faults was often "inferred from anomalous trends in mapped stratigraphical boundaries" and this source has been subject to major revision in recent years (cf. Fig. 2 with Dearman's Fig. 2 1963). An alternative explanation is that this zone is closely comparable to the Tamar Fault Zone, and that the discontinuities seen as events on the profiles may only have developed to the west of the fault. It seems apposite to note here that these events have been interpreted, by Day and Edwards (1983), as thrusts related to the emplacement of the Lizard Complex: Barnes *et al.* (1979) and Badham (1982) have suggested that this was tectonically transported northwards during the Devonian. The interpretation of the author would indicate that the fault zone was extant at this time.

The recognition of NNW-SSE and NW-SE trending fault zones of Hercynian age is of particular significance in this context, as their postulated interaction with E-W trending faults (Turner 1982b, in prep.) could permit variable block movements across the region and account for differences in both sedimentation and deformation timing. These fault zones could also account for the variable subsidence of blocks in the subsequent tensional phase of the Permo-Triassic.

Acknowledgements. I am very grateful to Dr. E. B. Selwood for his constructive criticism of an earlier draft of this manuscript and to Susan for her help during its preparation. I would also like to thank English China Clays, L.P.P.L.C. (Quarries Division) for access to Greystone Quarry and their boreholes in this area.

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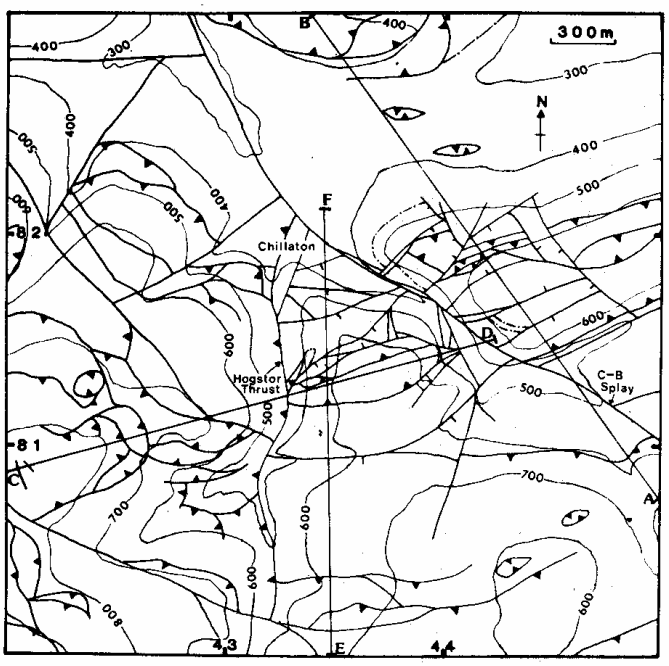


Figure 5a. Map of the Chillaton area showing geological boundaries

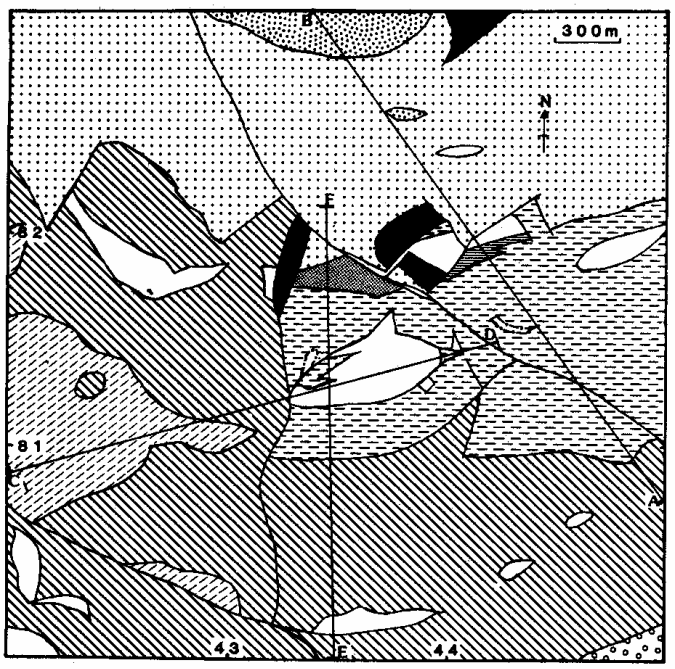
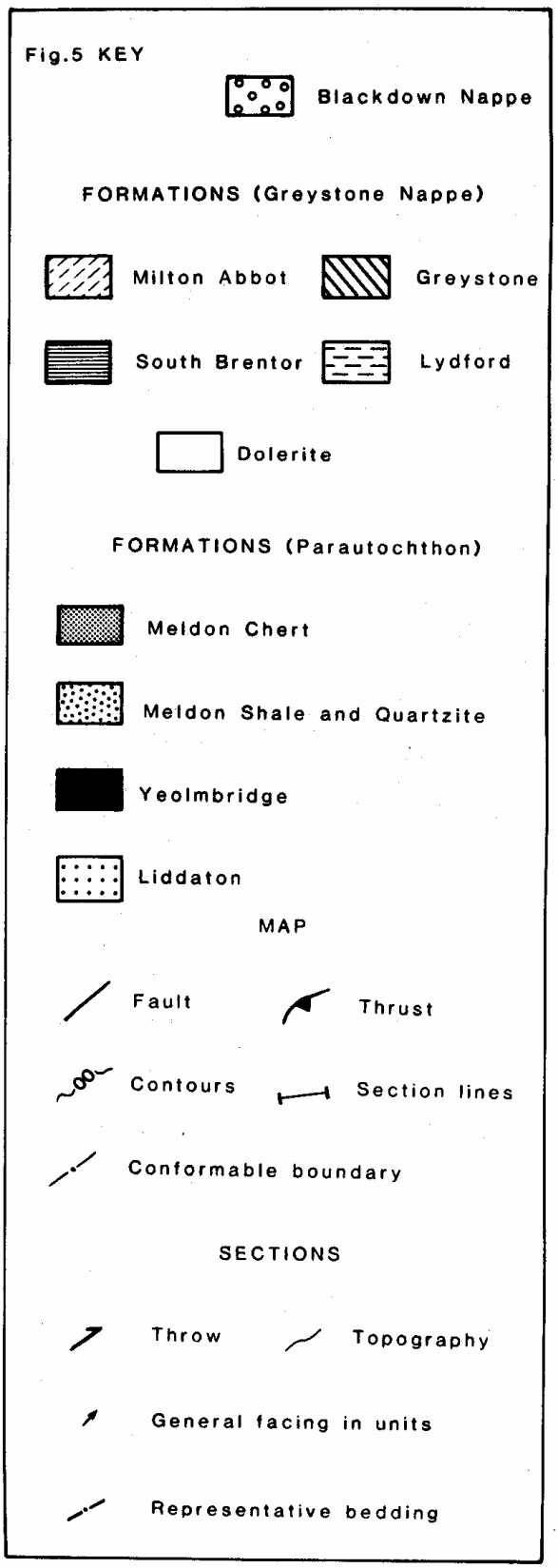


Figure 5b. Map showing formations recognized in the Chillaton area



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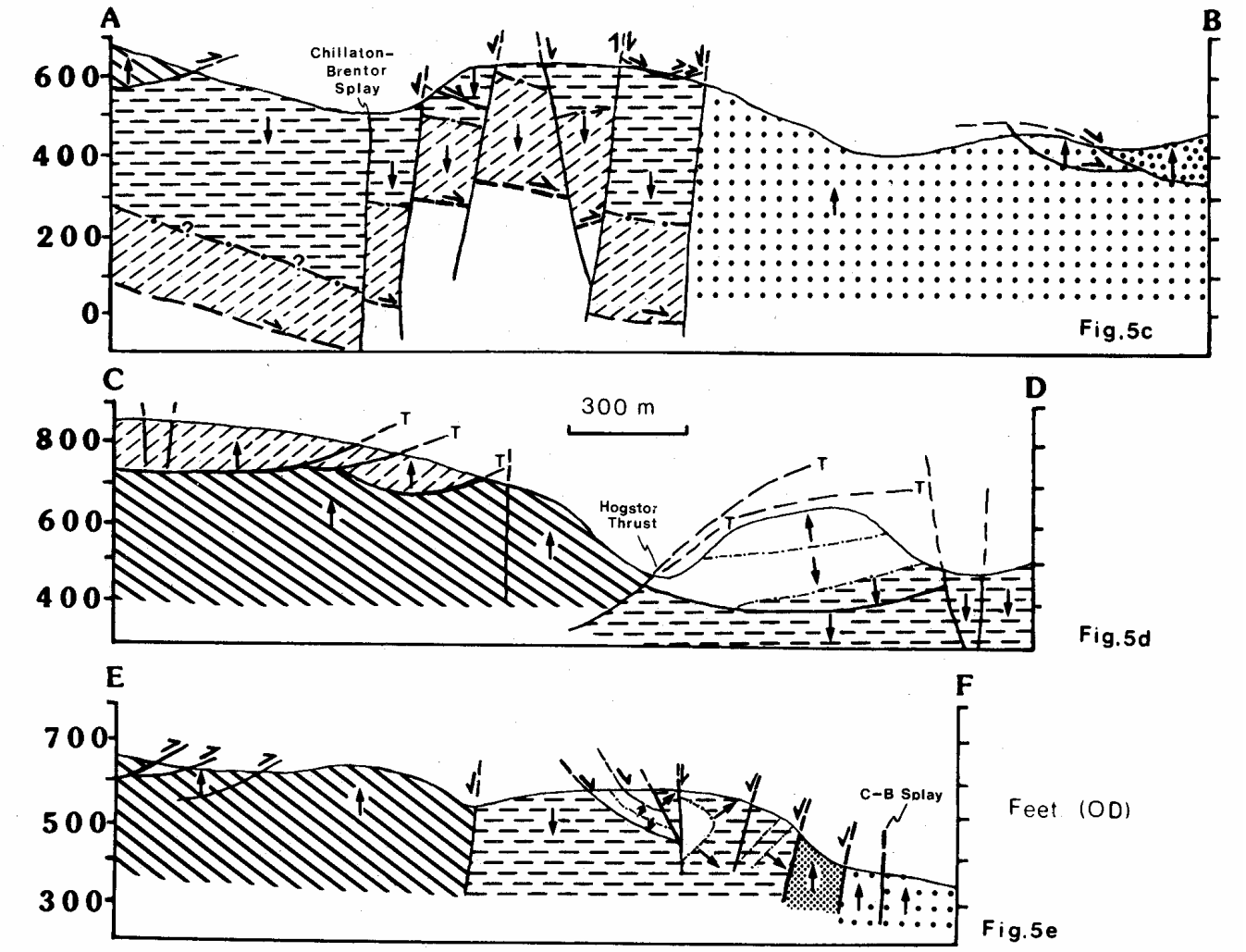

Figure 5c. -5e Cross sections in the Chillaton area (located on Figs. 5a and 5b)

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This paper is published with the approval of the Director, British Geological Survey (N.E.R.C.).

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Read at the Annual Conference of the Ussher Society, January 1984

The stratigraphy and structure of the Lower New Red Sandstone of the Exeter district

C. R. BRISTOW
R. C. SCRIVENER

Bristow, C. R. and Scrivener, R. G., 1984. The stratigraphy and structure of the Lower New Red Sandstone of the Exeter district. *Proceedings of the Ussher Society, 6,* 68-74.

Geological mapping of the area around Exeter has provided an opportunity to erect a revised lithostratigraphy for the Permian Strata taking account of previous classifications. A newly recognised, predominantly arenaceous, formation is described at the base of the sequence. The breccias that overlie these sands are divided using the influx of potassium feldspar fragments as a marker. A comparison is drawn between the sequence mapped in the southern part of the Crediton Trough, between Upton Pyne and Huxham, and that proved in the area south of Exeter with the stratigraphy of the intermediate area, around Pinhoe and Broadclyst, forming a link between the two districts. Lateral lithological changes in the lowest part of the sequence suggest that sedimentation may have been controlled by faulting.

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Introduction

Some 225 km sq of Exeter and its environs have been geologically surveyed during 1982 and 1983 by C. R. Bristow, R. A. Edwards, R. C. Scrivener and B. J. Williams of the British Geological Survey as a project commissioned and funded by the Department of the Environment (Fig. 1). The Permian outcrop of the area includes a small part of the eastern end of the Crediton Trough, a large area south of Exeter, and an area intermediate between these two east of Exeter. Separate Permian stratigraphies have been recognized and mapped in each of the first two areas; in the Pinhoe-Broadclyst area rapid lateral changes of facies obscure the stratigraphy. It is considered that faulting played an important part during sedimentation in the last area; rapid facies changes are ascribed to a response to syndepositional movement along faults resulting in the development of basins and highs in which local sequences were laid down. The transitional area may have been a topographic high throughout much of the early Permian times, and was not overtopped until late in the depositional history.

General stratigraphy and terminology

Ussher (1899; 1902) divided the Permian of the Exeter area into three broad division's which are, in ascending sequence: Breccia and Conglomerate, Lower Sandstone, and Lower Marls (Table 1). Each division was mapped over wide areas and allowed broad lithological correlations to be made. Detailed correlations, however, were not attempted in the absence of fossils. A prominent marker within the Breccia and Conglomerate, common to the Crediton Trough and the southern basin from Exeter to Dawlish, is the incoming of the potassium feldspar murchisonite. The influx of murchisonite as a datum was first recognised by Ormerod (1875) who named the strata in which it occurred the Murchisonite Beds. A number of local names have been used for these beds (conglomerates of Heavitree (De la Beche, 1839); Heavitree Conglomerate (Murchison, 1867); Exminster Conglomerates (Worth, 1890); Heavitree Breccia (Ussher, 1902), St. Cyres Beds (Hutchins, 1963); Kennford Breccias and Langstone Breccia (Laming, 1968); Exminster Breccias (Smith and others, 1974), and Coryton Breccias (Laming in: Durrance and Laming, 1982). It is proposed to use Heavitree Breccia Member (Bristow, 1984) for these murchisonite-rich beds.

Locally the Heavitree Breccia passes laterally over a very short distance into the argillaceous fine-grained sandy Monkerton Member (an unpublished field name used by Dr. D. J. C. Laming). In the area recently surveyed the Heavitree Breccia overlies the Alphington Breccia Member (Bristow, 1984) and these combined breccias are equivalent to the Teignmouth Breccias (renamed here the Teignmouth Breccia Formation) on the northern margin of the 1:50 000 Newton Abbot (339) Geological Sheet (1976). Beneath the Alphington Breccia is the locally developed, newly recognized, Whipton Formation (Scrivener, 1984) which rests on scattered outcrops of volcanic rocks, of the 'Exeter Traps' (Ussher, 1899, 1902) or the Exeter Volcanic Series (Dewey, 1935), or more usually on the Upper Carboniferous Crackington Formation.

South of Exeter the Heavitree Breccia is succeeded by a thick sequence of cross-bedded sandstones variously known as the Exminster Sandstone (Ussher, 1902), Dawlish Sandstones (Ussher, 1913), Dawlish Sands (Laming, 1966) or Clyst Sands (Laming, 1968). Of these names the slightly modified Dawlish Sandstone has been generally used (1:50 000 Newton Abbot (339) Geological Sheet, 1976), and it is proposed here to modify it further to Dawlish Sandstone Formation. In the Crediton Trough a thick sequence of alternating mudstones and sandstones has been recognised, and individually named and mapped, above the Heavitree Breccia. Initially (Bristow, 1983; Scrivener, 1983) these strata were named the Exeter Formation (Bristow, 1983), but from their stratigraphical position between the Heavitree Breccia and the Exmouth Mudstone and Sandstone (see below) it is clear that they are the stratigraphical equivalent of the Dawlish

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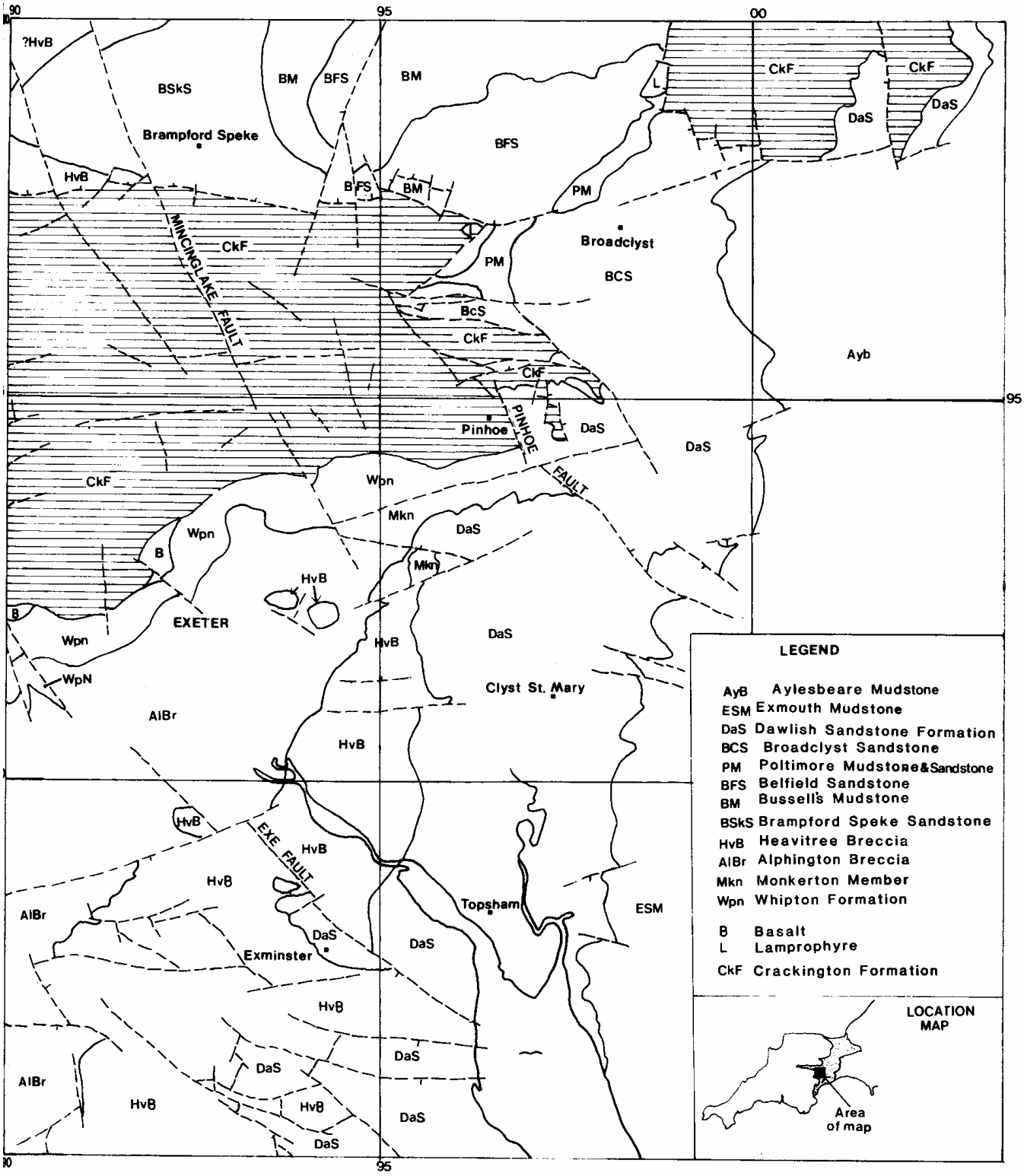

Figure 1 Geological sketch-map of the Exeter district

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Sandstone Formation and thus there is no necessity for retaining the name.

The Dawlish Sandstone Formation is overlain by a thick, dominantly mudstone, sequence--the Lower Marls of Ussher (1899); These were renamed the Exmouth Beds by Laming (1966) and subsequently the Aylesbeare Group (Smith and others, 1974). The name Exmouth Sandstone, was later (Laming, 1968) used in a more restricted sense for the lower part of the Lower Marls: the upper part being named the Littleham Beds (Laming, 1968), Littleham Formation (Henson, 1970) or Littleham Mudstones (Henson 1972). Further modifications to the name Exmouth Sandstone are as follows: Exmouth Formation (Henson, 1970: Smith and others, 1974; Bristow, 1983), Exmouth Sandstones and Mudstones (Henson 1972) and Exmouth Sandstone-and-Mudstone (Selwood and others, 1984). In this account this unit is referred to as the Exmouth Mudstone and Sandstone Member of the Aylesbeare Mudstone Formation.

The Exmouth Mudstone and Sandstone is only readily separated from the Littleham Mudstone Member (Edwards, 1984a) in the south of its outcrop where thick sandstones are present: the top of the uppermost one is taken as the top of the Exmouth Mudstone and Sandstone. Northwards the sandstones die out and the mudstones of the two formations merge into the undivided Aylesbeare Mudstone Formation (Edwards 1984b).

Initially Laming (1966; 1968) grouped the Dawlish and Exmouth formations as the Exe Group, hut this name appears to have fallen into disuse.

Detailed stratigraphy

Whipton Formation

This name has been given to the basal sequence of Permian strata in the Exeter district. The Formation comprises poorly sorted, commonly clay-rich, sandstones with breccias, mudstones and siltstones. The type section of the Whipton Formation is the stream section (SX944942) along the western boundary of the grounds of Northbrook School. Here beds of silty breccia and clayey sand rest with marked unconformity on a southerly-dipping surface of silicified and manganese-stained Crackington Formation. The basal breccia consists of about 0.3 m of coarse subangular cobbles of indurated sandstone and shale in a ferruginous cement. The sequence passes upwards through sands and breccias (about 9.0 m) into a predominantly argillaceous sequence (poorly exposed, c.40 m) which underlies the Alphington Breccia. Minor E-W faulting is common throughout the section.

The sandstones are generally very weakly cemented at outcrop and rapidly weather to a loose sand, red or reddish brown in colour and vary in grain size from coarse to silty or, in places, clayey fine sand. Some clay is commonly present, though relatively clean sands locally occur. The degree of sorting varies from poor to moderate. Bedding is typically parallel; cross-bedding is rarely developed and then only on a small scale. The breccias form the basal parts of sandstone units and commonly grade upwards into coarse sandstones. The thickness of the breccia units seldom exceeds 1.0 m and is commonly much less. Basal contacts are sharp and locally show channelling into the underlying finer sediment. Clast size is variable, with coarse breccias being subordinate to finer gritty units. Culm sandstone and slate predominate in the clast populations, with a very minor amount of quartz-porphyry and other igneous debris. In contrast the overlying A1phington Breccia is rich in igneous clasts. The breccia matrix may be red, rather coarse-grained, sand or brownish red, more argillaceous material. Except for the coarser units the breccias are matrix-supported. The red or purplish-brown mudstones and less common siltstones may form units up to several metres in thickness with prominent green and grey-green reduction zones. More usually, the argillaceous beds are relatively thin and sometimes present as discontinuous lenses within the sandstones.

The estimated thickness of the Whipton Formation in the type area is 100 m, but the formation thins towards the east and in the Exeter City Centre it is represented by only 20 m of red sand and mudstone.

Teignmouth Breccia Formation

Alphington Breccia Member. The oldest member of the Teignmouth Breccia Formation crops out south-west of Exeter in the Alphington to Shillingford area. It consists of gently dipping, clayey matrix breccias with little or no murchisonite feldspar fragments. The name Alphington Breccia was introduced (Smith and others, 1974) as part of the A1phington and Heavitree Breccias, but without definition, or subsequent amplification (Laming, in Durrance and Laming, 1982), except that it occurs at the base of the Permian. The Alphington Breccia is, in the southern part of the present district, equivalent to the lower part of the Teignmouth Breccia of the Newton Abbot (339) Sheet (1976).

The Alphington Breccia is poorly cemented, and this, combined with its high shale clast content which weathers to a clay, gives rise to more rounded, less steep, topography, than that of the Heavitree Breccia. Dips generally are less than 10 degrees and give rise to long dip-slopes. The outcrop is broken by large east-west strike faults. The maximum thickness is estimated to exceed 240 m. Because of the lack of cement the Alphington Breccia has not been quarried and there are few artificial exposures. Soils over the A1phington Breccia are generally clayey and contain fragments of Culm shale and sandstone, black chert-like rock, slate, lavas and a distinctive quartz porphyry. Where seen in roadside exposure the breccia fragments, mostly up to 4 cm, but some fragments 0.3 m in diameter, are set in a matrix of fine- to medium-grained sandstone; local thin seams (up to 0.1 m) of fine- and medium-grained sandstone occur. There are local occurrences of murchisonite feldspar and granitoid fragments, particularly in the upper part, but they form a very minor constituent of the total.

Heavitree Breccia. The type locality of the Heavitree Breccia is the quarry (SX950921) at Heavitree (also known as the East Wonford quarry in the earlier

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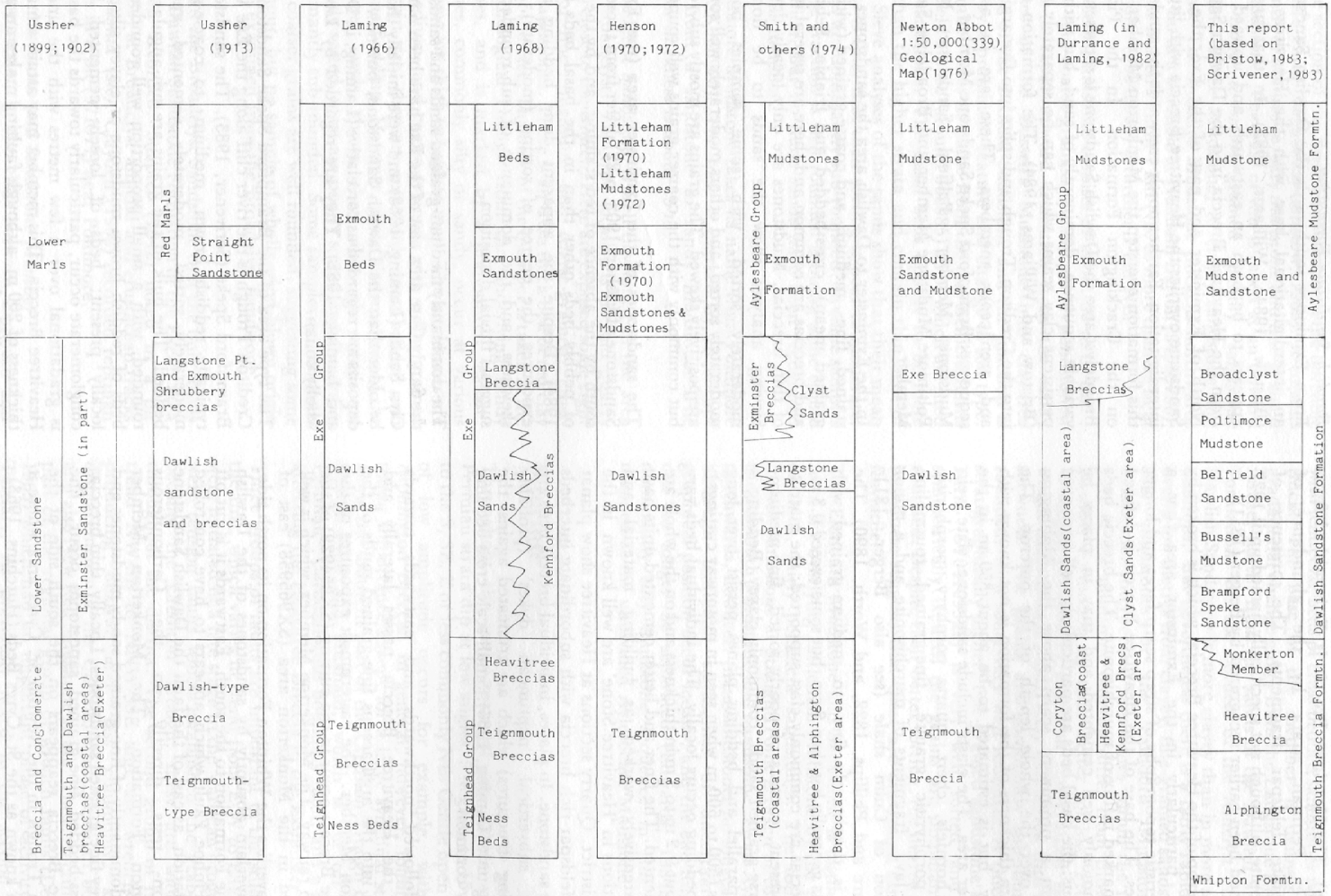

Table 1. Evolution of the stratigraphical terminology of the Exeter area

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literature). The Heavitree Breccia is distinguished from the older Permian strata by the abundance of murchisonite feldspar fragments. The presence of murchisonite, although not specifically named until 1827 (Levy), had been earlier recognized by Berger (1811).
The outcrop of the Heavitree Breccia, which is broken by faulting, particularly in the Exminster area, has a characteristic topography of steep scarp faces and longer dip slopes. The base of the Heavitree Breccia is defined by the marked influx of murchisonite. The breccia beds are commonly well cemented and may in places be calcareous; the basal beds are particularly well cemented and form a prominent scarp above the Alphington Breccia over the whole length of the outcrop. The maximum thickness of the formation is not known with certainty, but is estimated to be about 350 m in the Exminster area. The clasts include sandstone, slate, vein quartz, hornfels, chert, quartz porphyry, porphyritic spilite, porphyritic, granite, fine-grained tourmaline granite, lava, fragments of murchisonite and a minor' proportion of Culm shale (see also Berger, 1811; Conybeare and Phillips, 1822, and Worth, 1890). The breccia is dominantly fine- to medium-grained with fragments generally less than 8 cm, but some exceed 0.3 m. The breccias are commonly clast supported: the matrix, where present, consists of poorly sorted, fine-medium-and coarse-grained sands, commonly clayey. Bedding is usually parallel and bedding planes persist over long distances (500 to 600 m when seen in motorway cuttings): graded bedding occurs locally. The individual beds vary from about 0.2 m to 3 m in thickness and on the whole are well cemented. The finer and better cemented units have been extensively worked as building material and constructions in 'Heavitree Stone' are well known in the Exeter district. Quarry sections at Heavitree show planar beds of feldspar-rich breccia with subordinate interbeds of coarse sandstone. In places, individual breccia beds are cut by subvertical 'neptunean dykes' (probably dewatering structures) which are truncated against the overlying unit (Laming, 1966). Small-scale cross bedding is locally common.

When followed northwards and eastwards towards Pinhoe, the Heavitree Breccia passes laterally and upwards into the argillaceous fine-grained sands of the Monkerton Member. The easternmost exposures of the Heavitree Breccia occur in the cutting (SX954929) of the Exmouth railway. The Monkerton Member, which is well developed in the Monkerton area (SX965938) east of Exeter, overlies the Whipton Formation on its north side, and is overlain abruptly by sandstones of the Dawlish Sandstone Formation to the south. Eastwards it is cut off by the Pinhoe Fault which appears to have controlled sedimentation, as east of the fault the Dawlish Sandstone Formation rests directly on the Carboniferous Crackington Formation, The Monkerton Member consists dominantly of fine-grained sandy mudstones and very clayey fine-grained sandstones. Locally, thin breccias and cross-bedded fine-grained sandstones occur. The Heavitree Breccia reappears on the south side of the Crediton Trough in the Newton St. Cyres area where it has been known as the St. Cyres Beds (Hutchins, 1963).

Dawlish Sandstone Formation

From Exminster southwards the Dawlish Sandstone has an interdigitational base with the Heavitree Breccia (Bristow, 1984a, Williams, 1983). In that area there appears to be both an eastwards and upward passage from the Heavitree Breccia into the Dawlish Sandstone. North of Exminster and east of the Exe, the Dawlish Sandstone overlies the Heavitree Breccia with little or no interdigitation. Farther north-eastwards the sandstones of this formation overstep the Monkerton Member and rest on the Crackington Formation. In the Pinhoe to Broadclyst area the Dawlish Sandstone is represented by a variable sequence, dominantly of sand, but with several persistent clay seams which can be separately mapped (Bristow and Williams, 1984). The formation thickens into the Crediton Trough and splits into five argillaceous and arenaceous members. These are in ascending sequence: Brampford Speke Sandstone Member, Bussell's Mudstone Member, Belfield Sandstone Member, Poltimore Mudstone Member and Broadclyst Sandstone Member.

In the Exminster to Dawlish area the sandstones are red-stained, fine-, medium- and coarse-grained (with a fine-grained mean), cross-bedded and friable. Pebbly beds and breccias are common and there are several interbeds of thin breccias. Mudstones are only locally developed. Sieving analyses show the sands to be dominantly moderately sorted, with some samples poorly to moderately sorted, and others moderately well sorted. All are positively skewed. The grains are mostly sub-rounded, but commonly with the coarser grains well rounded.

The sandstones of the Exminster area (the Exminster Sandstone of Ussher, 1902, p.22) differ from those farther south by the absence of red staining, and by the absence of pebbly beds other than in the basal beds (Bristow 1984). Despite the apparent dune bedding in some exposures, the degree of sorting (moderate), positive skewness, and the dominance of sub-rounded grains suggests fluvial deposition.

The dominantly medium-grained sands and sandstones of the Sandy Gate area east of the Exe have been named the Clyst Sands (Laming, 1968) and were initially thought to be older than the Dawlish Sandstone, but now the two deposits are regarded as correlative (Laming, in Durrance and Laming, 1982). They are regarded by Laming as wholly eolian.

The type area of the Brampford Speke Sandstone in the Crediton Trough is the river cliff along the Exe Valley at Brampford Speke (Scrivener, 1983). The sandstones are typically reddish brown, medium- to coarse-grained, moderately sorted, moderately cemented and cross bedded. The bulk of the grains are sub-angular to sub-rounded, with a small proportion well rounded. Thin beds of reddish brown mudstone or clayey siltstone are locally present; beds of loosely cemented breccio-conglomerate occur, particularly towards the base which is gradational over a few metres with the underlying Heavitree Breccia. This member may attain a maximum thickness of 200 m, although faulting makes an accurate estimate difficult.

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The type section of the Bussell's Mudstone is the borehole at Bussell's Farm (95299873) (Bristow, 1983). It consists dominantly of mudstone, silty mudstone and fine-grained sandy mudstone. The base of this is taken at an upward change from a coarse-grained pebbly sandstone (Brampford Speke Sandstone) to clayey sandstone and mudstone. The Bussell's Mudstone thickens northwards from 24.4m in a borehole at Huxham Barton (94588788), to 50 m at Bussell's Farm I km NE, to more than 69 m at Columbjohn (95849964) a further I km NE.

The Belfield Sandstone consists of about 80 m of fine-, medium- and coarse-grained, locally pebbly sandstones. There is a rapid upward transition from the sandy mudstones of the Bussell's Mudstone into clay-free sandstone of the Belfield Sandstone and the junction is marked by a prominent feature and change in soil type. The member takes its name from the two Belfield farms and Belfield House (around 962985) all of which are sited on this sandstone. The basal beds appear to be more pebbly than the higher strata and contain thin beds of breccia. Sieve analyses of the sands show that their mean grain size is fine, that their sorting is poor to moderate; the grains are dominantly sub-rounded. In some samples the coarse fraction is well rounded showing some eolian contribution, but no dune sands have been recognized within this member.

The Poltimore Mudstone takes its name from Poltimore village (SX966970). The base is marked by an abrupt change from a sandstone to sandy mudstone. The maximum thickness is about 15 m,

The Broadclyst Sandstone is well developed in the area around Broadclyst. The type locality is the old quarry (983974) close to the church. Here are exposed some eight metres of fine-, medium- and coarse-grained, cross-bedded sandstones. There is an 8 cm thick bed of red clay in part of the face. The base of the member is taken at the sharp lithological change from the underlying sandy mudstones, and is marked by a pronounced feature. Mudstones, commonly silty or sandy, occur at various levels; some are thick enough to map. The sandstones are fine-, medium- and coarse-grained, with a fine-grained mean grain size; they are commonly cross-bedded and show dune bedding. However, grain size analyses of the 'dune'-type sands shows that although some are moderately sorted, most are poorly sorted and the grains are dominantly sub-rounded. Some samples are bimodal and the coarser grains are well rounded, suggesting some eolian input, but no sample of wholly wind-blown origin has been examined from the Broadclyst Sandstone.

It is within the Broadclyst Sandstone that the only fossils from the Dawlish Sandstone Formation of the Exeter area have been recorded. Shapter (1842) noted in a pit (SX989982) at Broadclyst annelid and crustacean tracks and *'Posidonia' (sic)*. Clayden (1908) recorded reptilian footprints in a pit (971971) near Poltimore.

Exmouth Mudstone and Sandstone Member

In its type section in the cliffs between Exmouth and Sandy Bay the Exmouth Mudstone and Sandstone consists of 255 m of red mudstones and silty mudstones with lenticular beds of cross-bedded sandstones up to 15 m thick. The base of the member is defined by the gradational change either from the Langstone Breccia, or farther north, from the sands of the Dawlish Sandstone Formation. On the borders of the Ashclyst Forest the Dawlish Sandstone is more clayey than in more southerly outcrops, but nevertheless there is a fairly rapid change over a few metres from clayey, fine-grained sands into the silty mudstones of the Exmouth Mudstone and Sandstone. Interbedded sandstones within the northern part of the mudstone Outcrop differ from those of the Dawlish Formation by their better sorting. From near Whimple (SY038970) a 0.3 m fine-grained sandstone within mudstones is moderately well sorted with negative skewness. The grains are dominantly sub-rounded to rounded, with some very well rounded. A thicker and stratigraphically higher, fine-grained sandstone at Strete Farm (SY04109522) is very well sorted and almost symmetrically skewed; the grains are dominantly sub-rounded. Despite their better sorting, the degree of grain roundness suggests that these are not eolian sands, but fluviatile.

The Exmouth Mudstone and Sandstone is distinguished from the overlying Littleham Mudstone by the presence of thick, fairly persistent sandstones. The top of the highest sandstone at Straight Point is taken as the top of the Exmouth Mudstone and Sandstone. From Aylesbeare northwards, however, these sandstones no longer persist, although lenticular sandstones of uncertain stratigraphical position occur locally, and the two formations can no longer be separated (Edwards, 1984b). The dominantly silty mudstone sequence in this area is mapped as the undivided Aylesbeare Mudstone.

Structure

Throughout the major part of the Exeter area the Permian strata dip, at low angles, § degrees to 15 degrees to the S, SE, E or less commonly NE. Gentle open folding of low amplitude occurs for example, within the Alphington Breccia. In general, however, folding is of minor Importance except for a large synclinal structure along the southern part of the Crediton Trough between Upton Pyne and Huxham, which is cut off by the Permian--Carboniferous boundary fault, and a syncline affecting the Heavitree Breccia in the Exminster area.

Three groups of faults are important in the study area: the first, which varies in trend from E-W to ESE-WNW were initiated in a late stage in the Variscan orogeny and were tension faults which established the graben that became the Crediton Trough; these are widespread throughout the district. The second group of fractures trend NW-SE to NNW-SSE and are wrench faults, producing dextral and, less commonly, sinistral, displacements of the Carboniferous/Permian boundary and also New Red Sandstone strata. These fractures form part of the regional group of wrench faults (Dearman, 1965) which had an important phase of activity in the Tertiary. As previously noted in the section on the Whipton Formation and Monkerton Member, the rapid change in sediment type across these faults in the study

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area suggests that they were active during, and partly responsible for the control of, lower New Red Sandstone sedimentation. This feature is particularly evident in the case of the Exe and Mincinglake faults (Fig. 1). The former is calculated to have a downthrow of approximately 150 m west. The third group of faults varies in trend from NNW-SSE through N-S to NNE-SSW and they are represented for example by a fracture with a westerly downthrow which passes to the east of Rewe and has a considerable downthrow to the west and is responsible for the presence of concealed Bussell's Mudstone beneath the Quaternary deposits of the Exe valley.

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Read at the Annual Conference of the Ussher Society, January 1984

A reinterpretation of the Meldon Anticline in the Belstone area

E. B. SELWOOD
J. M. THOMAS

Selwood, E. B. and Thomas J. M. 1984. A reinterpretation of the Meldon Anticline in the Belstone area. *Proceedings of the Ussher Society, 6,* 75-81.

The Lower Carboniferous stratigraphy of the Meldon Anticline near Okehampton is reassessed. Both autochthonous and allochthonous structural units and zones of fault rock are believed to be involved. The Meldon Anticline is reinterpreted as a late structure deforming a complex thrust pile during local underthrusting of the allochthon beneath the Crackington Formation of the autochthonous Culm Basin.

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Introduction

The Meldon Anticline, identified within the metamorphic aureole of the Dartmoor granite by Dearman (1959), has played a critical role in the interpretation of the Variscan structure of south-west England. Originally recognised in the extensive British Railways quarries at Meldon (SX571923) it was accepted as a reassuring reality in a region of complex and often disputed geology. However the isoclinally folded Meldon Slate with Lenticles Formation in the core of this moderately open fold has always appeared anomalous and evidence has now accumulated which suggests that the concept of this major fold should be reinterpreted.

A conformable succession of Upper Devonian to Namurian rocks has been mapped in the Meldon Anticline (Dearman 1959, Edmonds *et al.* 1968), a structure which was believed to form part of a continuum of folds showing progressive southward increase of southerly overturning from upright, north of Okehampton, to recumbent west of Dartmoor. (Dearman and Butcher, 1959 p.78). On the Tectonic Map of Great Britain (Dunning, 1966) it is represented as a major southward overturned structure with an arcuate trace round the northern edge of the Dartmoor granite. The westerly extension of this fold trace, which is interpretative, has been shown to be invalid, for it is projected west of Meldon through an area of recumbently folded strata, which is part of an allochthonous thrust pile derived from the south (Isaac *et al* 1982, 1983). The uppermost sheet of this allochthon, the Blackdown Nappe, is composed of Crackington Formation which, eastwards along the strike, is in lateral continuity with sandstones regarded as forming the southern overturned limb of the Meldon Anticline (Dearman and Butcher, 1959). Similarly, the Lower Carboniferous rocks comparable to those of the Greystone Nappe which underlie the Blackdown, Nappe, continue in unbroken outcrop eastwards to become involved in the anticlinal structure.

At a lower structural level, Isaac *et* al..(1983) have described in the well exposed section of Lydford Gorge, thick sequences of phyllonites and ultramylonites generated on the Main Thrust at the base of the Greystone Nappe. These rocks had previously been included in the River Lyd Slate with Lenticles Group (Dearman and Butcher, 1959), a unit equated with the Meldon Slate with Lenticles Formation. If this lithological correlation is valid, then the status of the Meldon Slate with Lenticles Formation must be questioned, and the interpretation of the Meldon Anticline, of which this unit forms the corer re-examined. These questions will be explored in this paper.

Since the costs levied for access to Meldon Quarry make extended geological investigation impractical, the authors have concentrated their effort some 5 km eastwards along the strike around Belstone (SX620937), particularly in the well exposed sections of the East Okement and Taw Rivers (Fig. 1). Here the structural pattern is said to be simpler (Dearman, 1962); a single anticline is recognised rather than the double anticline at Meldon.

The Succession

The stratigraphy currently employed by workers in the Okehampton area is derived from Edmonds *et al.*, (1968) and is summarised in Figure 2a. The present investigation suggests a division of this succession into an autochthon and two allochthonous units (Fig. 2b). The Crackington Formation adjacent to the granite is allochthonous and lithologically distinct from the autochthon north of Meldon, and it appears that the Meldon Chert Formation occurs in both the autochthon and allochthon.

Meldon Slate with Lenticles Formation

Exposure at the type locality in Meldon Quarry is currently indifferent, but the Meldon Slate with Lenticles Formation maintains its lithological character with remarkable uniformity throughout its outcrop, and excellent sections of this lowest level in the structure are available along the East Okement River and its western tributary, the Moor Brook (Fig. I and 3). It is difficult to find adequate descriptions of this formation in the literature but the impression is gained that it was considered that the distinctive lithology owed much to contact metamorphism by the Dartmoor granite. This is difficult to accept, for adjacent sedimentary rocks may only show light spotting.

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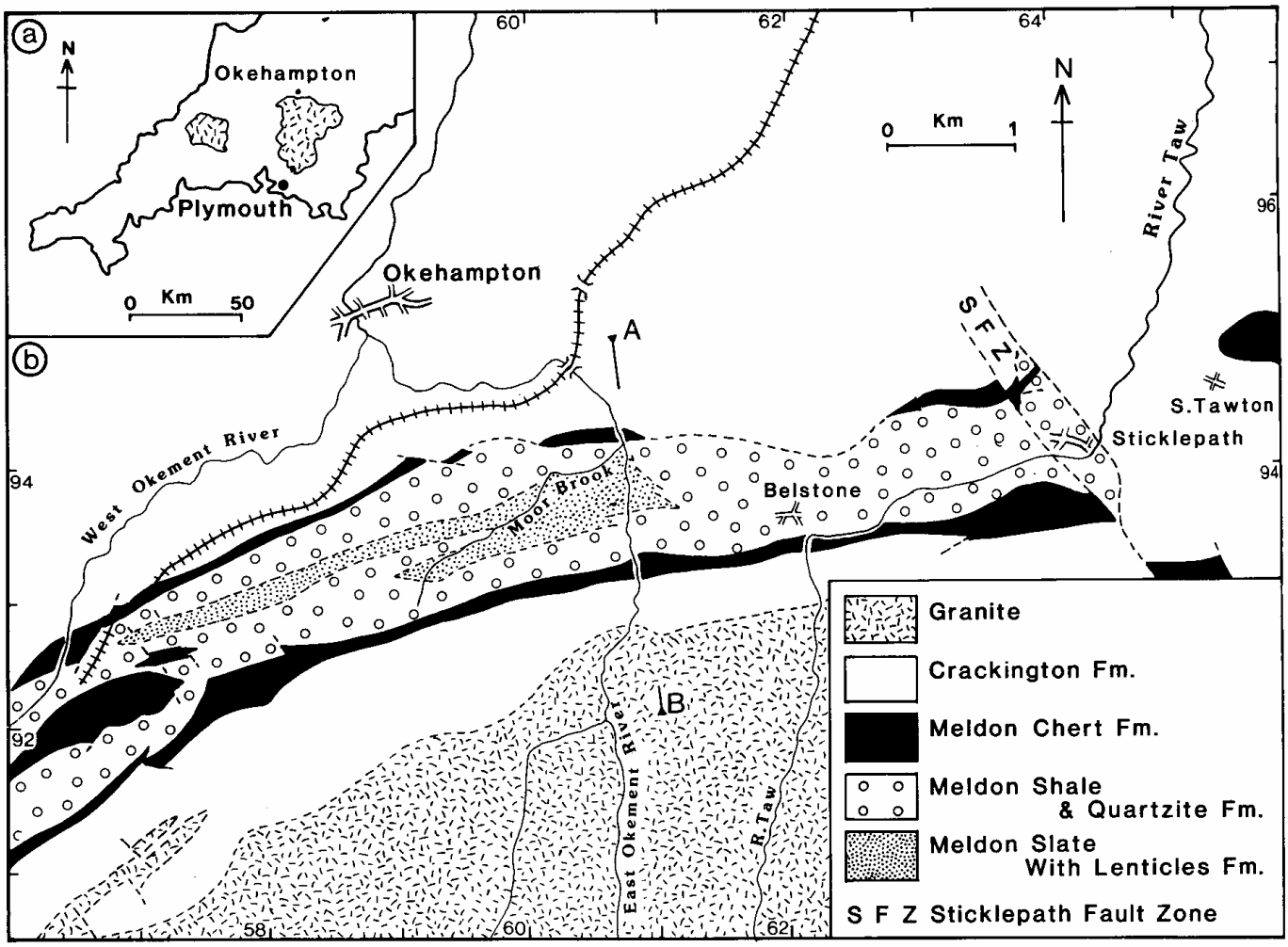

Figure 1. a Location Map. b The Meldon Anticline after BGS Sheet 324, dolerites and volcanics in Meldon Shales and Quartzites not shown, A-B, Line of Sections; Figure 3.

At outcrop the rocks are finely colour banded, pale to dark grey, on scales up to a few centimetres; they are even textured, highly siliceous and exceedingly fine grained. Conchoidal fractures develop evenly across colour bands and no preferred parting is observed. Totally siliceous 'lenticles' are represented but appear much less frequently than the formational name would suggest; they may lie either concordant to, or less frequently oblique to, the banding. Polished surfaces reveal the banding to be a tectonic fabric induced by progressive reduction of grain size, pervading both along and across a bedding lamination. For the most part this reconstitution is complete, but relict textures suggest that the original rocks comprised a fine sandstone, siltstone and slate succession. Individual thin colour bands may show internal ductile folding, with fold shapes approaching isoclines and extreme limb attenuation, and with the axial planes broadly parallel to the banding. Dr. K. P. Isaac having examined these rocks in thin section (pers comm.) reports the occurrence of textures associated with mylonitic deformation, but notes that the annealing textures present could be due to syndeformational recrystallisation, post deformational recovery or contact metamorphism. Lithological features thus allow the Meldon Slate with Lenticles Formation to be compared with fault rocks developed above the Main Thrust in Lydford Gorge, in strata which were formerly included in the River Lyd Slate with Lenticles Group (Dearman and Butcher, 1959). Here beds hosting the fault rocks are now (Isaac *et al.,* 1983) included in the Lydford Formation, a constituent member of the Greystone Nappe. A similar thrust-generated origin for the fault rocks of the Meldon Slate with Lenticles Formation, explains the previously problematic contrast of tectonic style between these rocks and the rest of the succession. It is not possible to make a definitive statement on the correlation of this thrust with the Main Thrust at Lydford.

Meldon Shale and Quartzite Formation

The Meldon Shale and Quartzite Formation (Edmonds *et al.,* 1968) consists of a dark grey slate interbedded with fine grained, often lenticular sandstones, cherts and a varied volcanic association of tuffs, agglomerates and lava. Dearman (1959) noted that the lowermost part of the formation in the type locality, Meldon Quarry, consists of 85 ft (26m) of black chiastolite slate with a few thin quartzite bands. Such slates, which form the

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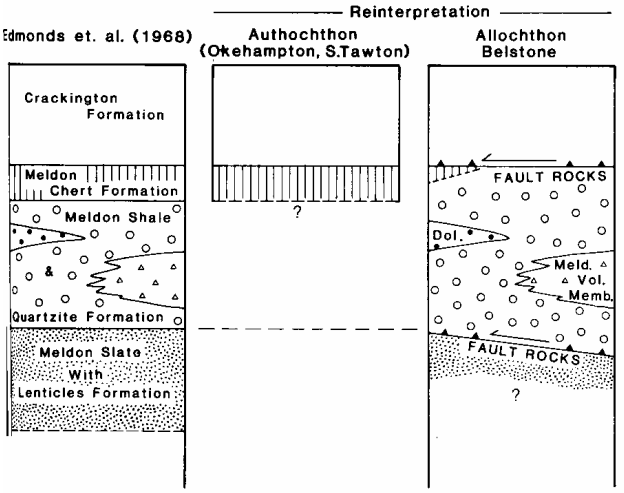

Figure 2. Comparison of currently accepted succession with the reinterpretation of this paper.

background lithology throughout the formation, are extensively developed at the lowest structural levels of the formation about Belstone but there they are associated with thicker, often tuffaceous sandstones, siliceous black slates grading into impure cherts, some thin agglomerates and dolerite sills.

Pyroclastic rocks, which occur more frequently in the upper parts of the succession, range from fine crystal and lithic tuffs to coarse agglomerates, and are usually associated with hard black siliceous shales locally developing into impure cherts. Particularly distinctive is a pyroclastic breccia in which siliceous sedimentary and subordinate volcanic clasts are set in a fine grained basaltic matrix. Agglomerates of this type are spectacularly displayed east of Belstone, in crags above the north bank of the River Taw (SX625936) and forming Ivy Tor (SX628936). The close packing of clasts is variable, and a poorly defined lenticular pattern of clast size variation can be recognised. In parts of the unit lacking lithic clasts, the angular feldspar fragments in a basaltic matrix give a superficial field appearance of a dolerite. This rock type appears to have been formed by lava debris flows.

Dolerites map as sill-like bodies throughout the formation, and we can find no evidence to suggest that they postdate deformation. Although they are highly altered most retain a coarse gabbroic appearance with ophitic textures. Contact metamorphism has undoubtedly contributed to the present mineralogy (Edmonds *et al.*, 1968) but the low grade (chiastolite slate) of the country rock suggests that much of the alteration and texture is primary. A spilitic association, comparable to that of the albite dolerites of adjoining areas (Chesher, 1969, Turner, 1982a) appears possible. Such an interpretation would be consistent with field relations where an intimate mixing of sediment and dolerites can be demonstrated. The outcrop pattern of the dolerites represented on the 1:63,360 IGS map (Sheet 324) appears to be a broad generalisation through poorly exposed ground, which may grossly exaggerate the thickness and continuity of individual sills.

In the lowest observed levels of the formation in the East Okement and Moor Brook sections, the gently north dipping junction with the Meldon Slate with Lenticles Formation is locally gradational with the lowest 5m of black slate of the Meldon Shale and Quartzite Formation. These slates develop occasional thin lenticular zones of grey siliceous fault rock which irregularly become more frequent, downward, until the lithology is wholly that of the underlying formation. For some 20rn above this zone large pods of agglomerate, dolerite, sandstone and siliceous slate occur irregularly in a highly sheared zone. Where stratigraphic horizons can be identified within the formation, as with the siliceous slates of West Cleeve (SX609941) and their overlying agglomerates, they are first disrupted and then truncated as they enter this zone. A tectonic mélange appears to be represented which is associated with the fault rocks of the underlying Meldon Slate with Lenticles Formation. Both appear to have been generated by major thrusting and it follows that the Meldon Shale and Quartzite Formation should be regarded as allochthonous.

Meldon Chert Formation

At its type section in the syncline in the axial zone of the Meldon Anticline in Meldon Quarry, the Meldon Chert Formation is described (Edmonds *et al.*, 1968) as 73m of interbedded dark grey slates, siliceous mudstones, radiolarian cherts and impure black limestones showing normal stratigraphic contacts with the underlying Meldon Shale and Quartzite Formation. From evidence presented above, this conformity between the formations means that both occur in a structurally allochthonous unit. No higher parts of the succession are recorded in the quarry sections but in inliers to the north (e.g. South Tawton, SX658951) and east (e.g. Drewsteignton, SX731911) the Meldon Chert Formation is reported (Edmonds *et al.*, 1968) to be conformable with autochthonous Crackington Formation of the main 'Culm Basin' of Devon. This formation thus appears in allochthonous and autochthonous successions. Broadly similar chert and shale sequences are represented in the highest parts of the Lower Carboniferous throughout south-west England. In west Devon and north Cornwall cherts belonging to the allochthon are referred to the Firebeacon Chert Formation. It would be logical to consider extending this usage to the allochthonous cherts occurring in the Meldon area.

In the East Okement section, Crackington Formation represented in the northern, "normal" limb of the Meldon Anticline is held (Dearman, 1972, Edmonds *et al.*, 1968) to succeed conformably cherts referred to the Meldon Chert Formation. However, greywackes of the former formation are clearly inverted and the cherts are limited to the immediate area of the valley floor beneath a gently north-dipping fault which introduces the Crackington Formation (SX607942). Although 'chert' is an appropriate field term for this tough, colour banded (pale grey-white) and often very finely laminated very fine grained siliceous rock, Dr. K. P. Isaac notes that in thin section these rocks show classic fault rock textures. One section in particular included all of the typical features of quartzo-feldspathic mylonite namely, deformation lamellae, subgraining, new grain growth resulting in a

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reduction of grain size by a factor of about 10, ribbon texture and a fine grained foliated matrix containing porphyroclasts. The intensity of deformation is extremely heterogeneous and fine grained ultramylonitic bands alternate with blastomylonitic areas which show a degree of syn- or post-deformational recrystallisation. Other sections showing similar mylonitic and cataclastic textures also revealed the effects of contact metamorphism because of their different mineralogy. Immediately upstream similar 'cherts' form a siliceous matrix welding together a range of large clasts including dolerite, tuff, agglomerate and sandstone which is interpreted here as a tectonic mélange.

On the southern limb of the Meldon Anticline, the Meldon Chert Formation has been mapped previously as a continuous horizon which crosses the East Okement River above Chapel Ford (SX608952). Present investigations around Belstone indicate that these 'cherts' occur in large (tens of metres long) isolate pods at the contact between the Meldon Shale and Quartzite and the southern tract of Crackington Formation. In hand sample, many of these rocks are difficult to distinguish from the 'cherts' in the northern part of the East Okement section which are now recognised as fault rock. Variation in texture in these 'cherts' reflects differences in host rock lithology. About Ivytor Mine (SX628936) where the enclosing rocks constitute a black slates with sandstone sequence, the 'cherts' have a background sandstone texture which in hand sample merges into and is interleaved with pale, very fine grained siliceous material. Comparable relationships on a larger scale can be determined in the upper reaches of the East Okement River, where metamorphosed sandstones develop progressive 'chertification' as they approach the mapped outcrop of the Meldon Chert Formation (SX608932). Hand samples from these cherts show relict sandstone textures in beds interleaved with fine grained 'chert'. In the floor of the Taw Valley (SX622934) siliceous rocks have developed within a black slate lithology, whilst large loose blocks of 'chert' on the north east side of Watcher Hill (SX614932) appear to be hosted in agglomerate. It is suggested that all of these 'cherts' have a common fault rock origin.

Farther west, along the southern margin of the Meldon Anticline and south of Meldon Quarry, similar fault rocks may occur, but it appears the cherts of sedimentary origin may also be involved in the fault zone. Here the substantial thickness of Meldon Chert Formation mapped by Dearman (1959), in conformable contact with the Meldon Shale and Quartzite Formation and would be interpreted as part of the allochthonous unit comparable with those of the axial zone of the anticline.

Crackington Formation

The Namurian to Lower Westphalian Crackington Formation (Edmonds *et al.*, 1968) consists of a thick flyschoid sequence of interbedded dark grey slates and fine grained greywackes showing the sedimentary structures of a distal turbidite facies with an E-W sediment supply. In the Okehampton area, two principal outcrop tracts occur (Dearman 1959, Dearman and Butcher, 1959 and IGS Sheet 324), that adjacent to the Dartmoor granite being separated from the main outcrop in the 'Culm trough' of south-west England by the Lower Carboniferous rocks of the Meldon Anticline. Since a conformable relationship between these rocks was previously assumed, the pattern of outcrop supported the concept of a major anticlinal fold. However, in the Belstone area sole markings show that the Crackington Formation forming the 'normal' northern limb of the Meldon Anticline is inverted and overthrusts the Lower Carboniferous succession on a gently north dipping fault.

Sandstones in the Crackington Formation adjoining the granite which are ascribed to the inverted limb of the Meldon Anticline, are too poorly exposed and recrystallised to allow younging directions to be determined. These too lie in faulted contact with the Lower Carboniferous rocks. In contrast to the greywackes of the main basin, these sandstones appear to show a higher sandstone to shale ratio and a lower clay content. Such lithologies are locally developed in the Crackington Formation of the Blackdown Nappe where they appear to represent southerly derived flyschoid sediments of more proximal character than the greywackes of the main basin. Since there is continuity of outcrop between the Crackington Formation adjacent to 'the granite and that of the Blackdown Nappe, it can be argued that both are represented in the same structure.

Summary

From the stratigraphic evidence presented above, and shown in figure 2, a number of important points emerge which are of relevance to the structural discussion which follows, namely:-

1. The primary lithology of the Meldon Slate with Lenticles Formation has been largely obscured' by the generation of fault rock associated with major thrusting.

2. The lithologies of the Meldon Shale and Quartzite Formation are diverse and show a high degree of lenticularity much of which derived from tectonic dismemberment.

3. Rocks formerly mapped as Meldon Chert Formation include fault rocks generated by major thrusting, sedimentary cherts conformable with the autochthonous Crackington Formation of the main flysch basin of south-west England and sedimentary cherts conformable with the Meldon Shale and Quartzite Formation of the allochthon.

4. Two tectonically separated facies of Crackington Formation are recognised.

5. Fault rocks associated with major thrusts have been identified at two principal levels; one separating the Meldon Slate with Lenticles Formation from the Meldon Shale and Quartzite and the other separating the Meldon Shale and Quartzite Formation from Crackington Formation adjacent to the Dartmoor granite. These thrusts divide the succession into an autochthon and two allochthonous units. (Fig. 4). Lithologically the Crackington Formation adjacent to the Dartmoor granite is similar to and continuous with the Crackington Formation of the Blackdown Nappe whilst the Meldon Shale and Quartzite Formation with its conformable chert

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formation could either represent part of the Greystone Nappe or Parautochthon disrupted in advance of that nappe.

Structural Interpretation

Hitherto it has been assumed that the Meldon Anticline was based upon a sound stratigraphy, to the extent that the repetition of formations provided the strongest supportive evidence for its existence. The stratigraphic evidence now presented, profoundly weakens the anticlinal concept. In the Belstone area, no conformable contacts are observed. The Crackington Formation previously referred to the northern, *"normal"* right way up limb is inverted and of contrasting facies to that of the southern limb; the Meldon Chert Formation in both limbs is not a stratigraphic formation but fault rock as is the Meldon Slate with Lenticles Formation of the anticlinal core. The consequence is that stratal dips alone define the Meldon Anticline; but even these have to be interpreted with care. In the East Okement section, although Dearman and Butcher's (1959, Fig. 10) observations on the dip of the Meldon Slate with Lenticles Formation and Meldon Shale and Quartzite Formation contact which define the core of the Meldon Anticline are valid, evidence now presented indicates that this junction represents a thrust. Similarly the steeply inclined contact between the southern tract of Crackington Formation and the Meldon Shale and Quartzite Formation is interpreted as a thrust rather than a conformable contact in the inverted limb of the fold. Both thrusts appear to be folded concordantly into the structure known as the "Meldon Anticline". This concordancy argues for a second deformation, rather than thrust ramping. Dearman's (1972) observations are reinterpreted accordingly (Fig. 3a and b).

The high degree of stratal disruption in the Meldon Shale and Quartzite Formation, the lack of symmetry of lithologies within the formation either side of the fold axis and the lenticularity of the igneous bodies within the succession can now be explained in terms of an early deformation associated with the emplacement of the nappes This early deformation affords ready explanation for the hitherto problematical small scale tight recumbent folds which can be observed from time to time within the formation at Meldon Quarry (e.g. Dearman

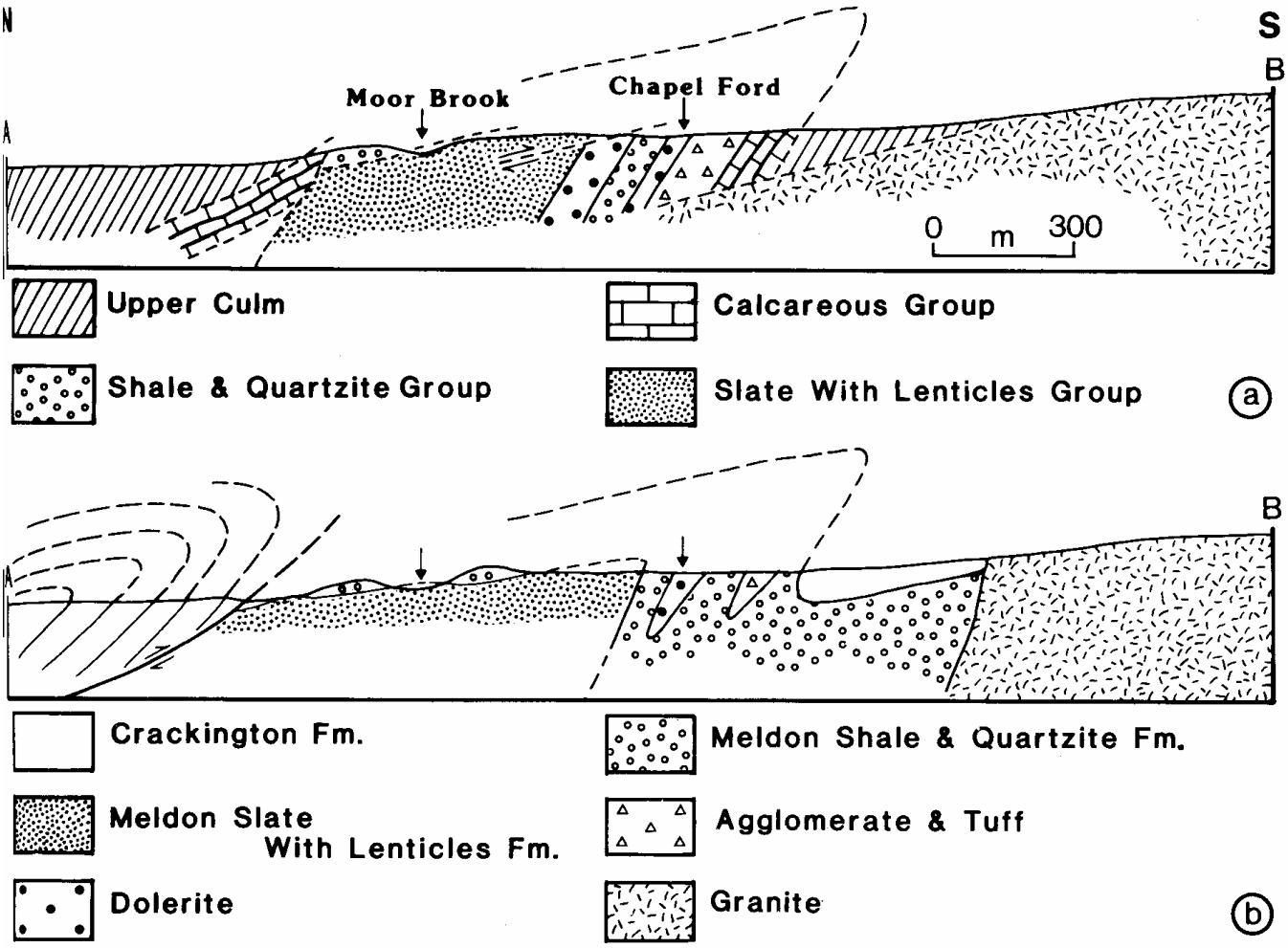

Figure 3. East Okement Section comparing existing interpretation (a) of Dearman and Butcher 1959. Dearman 1962, with (b) reinterpretation of this paper.

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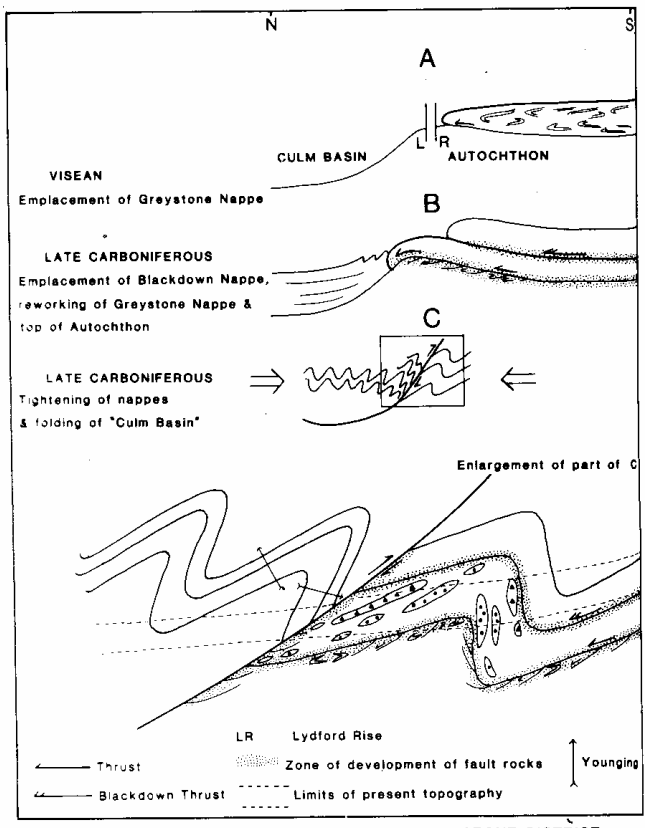
1959, Fig. 9b, c) as well as those in the Meldon Slate with Lenticles.

Evidence has been presented which suggests that the allochthonous strata represented in the Meldon Anticline belong to the Blackdown Nappe and either the Greystone Nappe or a parautochthonous unit which is developed extensively farther west. Observations in the Belstone area now make clear the relation of these nappes to the flysch basin to the north. Here allochthonous strata are overridden, on a low angle north dipping thrust, generating a tectonic m61ange, by Crackington Formation of the main flysch basin (Fig. 4). From north of Okehampton these beds are involved in folds which southwards become progressively more overturned to the south (Fig. 4). The development of these folds is linked to the generation of the second folds of similar geometry affecting the underlying thrust units. This may be associated with the late Westphalian compression of the Culm basin which produced the main folding in central and north Devon. Such movements would undoubtedly have been accompanied by late tightening of the nappe structures and could well have produced local underthrusting beneath the Crackington Formation with the generation of local southward facing backfolding in this superficial cover. The relative importance of over-and under-thrusting is unknown, but the Meldon Anticline appears to have formed within the imbricate zone of this thrusting beneath the Crackington Formation. The importance of north dipping faults in Meldon Quarry reported by Dearman (1959) would be consistent with this view.

The Meldon structures are not continued eastwards beyond the Sticklepath Fault, where conformable Vis6an and Namurian successions are believed to be autochthonous. This fault is part of the major north-west --south-east fault system of south-west England now recognised as deep seated structures which elsewhere (Turner 1984) have played a critical role in delimiting nappe structures.

The recognition of a northward directed thrust and nappe terrain between Bodmin Moor and Dartmoor (Isaac *et al.* 1982, 1983, Turner 1982b) generated a "confrontation" problem with the original concept of southward transport which included the Meldon Anticline. In the reinterpretation presented here a major southward directed structure is neither necessary nor justified, and the southward verging folds in the Crackington Formation north of Meldon are interpreted as local backfolds developed in response to underthrusting. Comparable structures are developed to the west along the strike on the southern margin of the Culm basin on the north Cornwall coast, (Selwood, Stewart and Thomas, in press, Selwood and Thomas, 1984), though locally between these areas (Turner, 1982a) northward transported units overthrust the Crackington Formation of the Culm Basin.

Acknowledgements. The authors would like to thank Mrs. V. Ellis for typing the manuscript and Mr. J. S. Jones for drafting the diagrams, and the University of Exeter for funding the field work.



SCHEMATIC STRUCTURAL HISTORY OF THE BELSTONE DISTRICT

Figure 4. A model suggesting the structural history of the Belstone district, in which the Meldon Shale and Quartzite Formation is regarded as derived from the Greystone Nappe (Isaac *et. al.,* 1982).

A. Emplacement of Greystone Nappe by gravity sliding in the Late Visean, possibly separated from the Culm Basin by the Lydford Rise (Isaac *et. al.,* 1983) of unknown width.

B. Late Carboniferous arrival of Blackdown Nappe which reworks the Greystone Nappe and the top of the autochthon, causing some underthrusting beneath the Culm Basin.

If the Meldon Shale and Quartzite Formation is par autc)(hthonous then Stage A may be eliminated.

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Read at the Annual Conference of the Ussher Society, January 1984

The North Curry Sandstone Member (late Triassic) near Taunton, Somerset.

G. WARRINGTON
B. J. WILLIAMS

Warrington, G. and Williams, B,J. 1984. The North Curry Sandstone Member (late Triassic) near Taunton, Somerset. *Proceedings of the Ussher Society, 6,* 82-87.

The North Curry Sandstone Member occurs in the upper part of the Triassic Mercia Mudstone Group succession in Somerset and crops out principally to the east of Taunton, at and near the type locality, here designated, near North Curry. The member varies in thickness from about 2.5 m to at least 7.5 m; the thicker developments consist largely of sandstones and pass laterally into thinner sequences comprising siltstones and fine-grained sandstones with interbedded grey and green bioturbated mudstones. The member has yielded palynomorphs of late Triassic (Carnian) age and a sparse macrofauna with both terrestrial and subaqueous, probably largely euryhaline, components. It is analogous in facies to the Schilfsandstein in the late Triassic succession in Germany and is interpreted as the deposit of a deltaic or estuarine environment of mudflats traversed by channels. The North Curry Sandstone Member is a correlative of the Weston Mouth Sandstone Member in the Mercia Mudstone Group of Devon and of the Arden Sandstone and Dane Hills Sandstone members in that group in, respectively, Worcestershire and Warwickshire, and Leicestershire.

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Introduction

The name North Curry Sandstone Member, hereinafter NCSM, was introduced by Warrington et al. (1980, p.60) for an arenaceous unit in the upper part of the Mercia Mudstone Group succession of the Taunton district which had previously been termed "Upper Keuper Sandstone" (Ussher 1908). The member occurs in a presently unnamed mudstone formation which lies between the Somerset Halite Formation, proved in boreholes at Puriton (McMurtie 1912) and Burton Row, Brent Knoll (Whittaker and Green 1983), and the Blue Anchor Formation, the highest formation in the Group. It varies in thickness from about 2.5 m to at least 7.5 m, and comprises grey and green mudstones and siltstones with locally thick beds of white to pale brown sandstones and in these respects, and in its relatively fossiliferous nature, contrasts markedly with the predominantly reddish brown mudstones that constitute the greater part of the Mercia Mudstone Group succession.

The principal outcrop of .the NCSM occurs in the type area of the unit to the east of Taunton and runs approximately east-north-eastwards from the vicinity of Knapp (ST300255) to near Stathe (ST374290), a distance of about 9km. All but the most easterly part of this area (Fig. 1) was mapped by one of us (BJ.W.) between 1977 and 1979 during the resurvey of the Taunton district (1:50 000 Geological Sheet 295; Edmonds and Willams in press). Impersistent sandstones have been reported from comparable horizons in the Mercia Mudstone Group elsewhere in that district (Moore 1867 1881; Ussher 1908) and to the south, in the adjoining Wellington district (Geological Sheet 311; Ussher 1906). However, some reported occurrences, for example that observed by Moore (1867, 1880, 1881) in an excavation for the foundations of a house at Ruishton (ST265250) near Taunton, were not traced during the recent survey. The lithostratigraphy of the NCSM at the type locality, here designated, near North Curry, and at other sites in the type area, is documented on the basis of work carried out during that survey. The results of a palynological study carried out by one of us (G.W.) on the member, and contiguous beds in the Mercia Mudstone Group, in its type area, and a review of macrofossils recorded from the unit in the Taunton district, are presented.

The North Curry Sandstone Member in the type area

The principal outcrop of the NCSM (Fig. 1) is on an east-north-east trending ridge formed of Mercia Mudstone Group deposits and flanked, to the north and south respectively, by alluvium of the River Tone and West Sedge Moor. The beds cropping out in this ridge dip at 5° to 10° in a south-easterly direction and the topographic feature results from the presence of the comparatively resistant NCSM. Exposures of the NCSM and contiguous beds are, however, scarce and the outcrop of the member is defined largely by a topographic feature comprising a steep northward-facing scarp and a gentle southward-facing dip slope.

The most westerly manifestation of the NCSM in the type area is a weak feature seen on a hillside to the west and north of New Barn (Fig. 1). In a water borehole (ST29632447) at New Barn 5.49 m of grey marl resting on 3.35 m of grey sandstone were recorded overlying grey and red marls (driller's records); the grey sandstone and some, at least, of the contiguous grey marls here are thought to represent the NCSM. About 0.5 km to the north, the feature is traceable around Knapp Hill and sandstone is present in laneside exposures (ST29752520) on the south side of the hill. The outcrop at Knapp is separated at the line of a north-south-trending fault from a more pronounced north-facing scarp feature which

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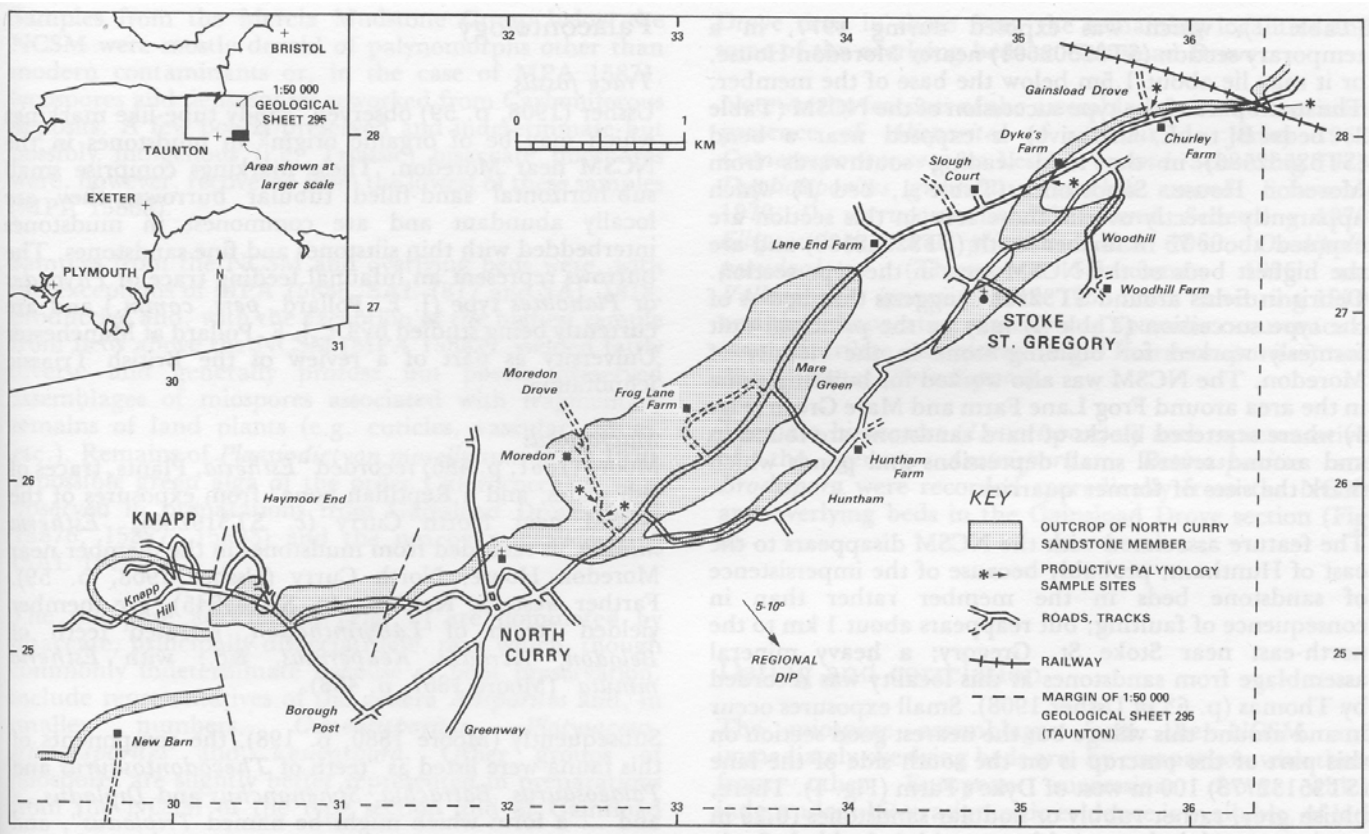

Figure 1. Location map and outcrop of the North Curry Sandstone Member in the type area; based, with permission, upon British Geological Survey 1:50,000 Geological Sheet 295 (Taunton)

extends eastwards for about 1 km towards North Curry and terminates at the line of a north-north-east-trending fault; the site of an old building-stone quarry (ST30502536) on this part of the outcrop is now overgrown. About 1 km to the south, near Borough Post, 25m of red, green and grey marl resting on 1.22m of red and grey sandstone on 3.35m of greenish grey sandstone overlying red marl were proved in a borehole (STB0552430); the 4.57m of sandstone recorded here is interpreted as the NCSM. To the east of the north-north-east-trending fault crossing Knapp Lane, the outcrop of the NCSM is narrow and difficult to trace precisely, but to the north-east of North Curry it widens considerably and forms a broad plateau extending from Moredon to Mare Green with a minor branch extending to Huntham. The best extant exposures of the member occur in this part of the outcrop, and those in the private road leading north-north-west from the North Curry to Stoke St. Gregory road, past Moredon House, to Moredon Drove constitute its type section (Table 1); the base of the unit may be visible in this section but its upper boundary is not seen.

Beds in the Mercia Mudstone Group beneath the NCSM crop out on the northward-facing scarp capped by the member, and are exposed in Moredon Drove where about 15.5m of red brown blocky and sandy mudstones are overlain by a greenish grey arenaceous unit comprising up to 0.5m of hard nodular sandstone overlain by 1.0m of flaggy sandstone. This unit, exposed at ST32452626, may be equivalent to bed A, the basal bed of the NCSM

TABLE 1. Succession exposed in the type section of the North Curry Sandstone Member near Moredon House (ST32502601 to 32552580)

F. Sandstone:	brownish grey, thin to medium bedded, with thin brownish grey sandstone beds and laminae (top and base not seen) (gap in section)	1.55m seen
E. Sandstone:	greenish grey, fine-grained, finely micaceous; with greenish grey mudstone intercalations and some bioturbation	2.00m seen
D. Mudstone:	greenish grey, soft, sandy; some bioturbation and sand-filled burrows	0.14m
C. Sandstone:	greenish grey, fine-grained, finely micaceous; cross-bedded in parts	0.36-0.55m
B. Sandstone:	brown, fine-grained, finely micaceous; with thin greyish brown sandy mudstone beds and thin bluish green mudstones with mud-cracks. (gap in section)	1.42m seen
A. Sandstone:	Sandstone: greenish grey, fine-grained, micaceous, cross-bedded; with thin green mudstone beds and laminae; (seen in temporary exposure, top and base not seen).	1.50m seen

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(Table 1), which was exposed during 1977, in a temporary section (ST32502601) nearer Moredon House, or it may lie about 1.5m below the base of the member.

The main part of the type succession of the NCSM (Table 1, beds B to E, inclusive) is exposed near a bend (ST32532586) in the road leading southwards from Moredon House. Sandstones (Table 1, bed F) which apparently directly overlie those seen in this section are exposed about 75 m farther south (ST32552580) and are the highest beds of the NCSM seen in the type section. Debris in fields around ST325260 suggests that bed A of the type succession (Table 1) may be the principal unit formerly worked for building stone in the vicinity of Moredon. The NGSM was also worked for building stone in the area around Frog Lane Farm and Mare Green (Fig. 1) where scattered blocks of hard sandstone are found in and around several small depressions and ponds which mark the sites of former quarries.

The feature associated with the NCSM disappears to the east of Huntham, probably because of the impersistence of sandstone beds in the member rather than in consequence of faulting, but reappears about I km to the north-east near Stoke St. Gregory; a heavy mineral assemblage from sandstones at this locality was recorded by Thomas (p. 63 in Ussher 1908). Small exposures occur in and around this village but the nearest good section on this part of the outcrop is on the south side of the lane (ST35132773) 100 m west of Dyke's Farm (Fig. 1). There, bluish grey, rather rubbly or nodular sandstones (0.25 m seen) are overlain by greenish grey, thinly bedded, finely micaceous sandstones with thinner greenish grey mudstone beds and lenses, some with bioturbation, (1.03m seen).

About 0.7 km farther north-eastwards, in Gainsload Drove (ST35692814), the NCSM is represented by about 2.5 m of greenish grey, finely micaceous, thinly bedded sandstones interbedded with greenish grey mudstones, some of which are bioturbated. Overlying beds in the Mercia Mudstone Group are also visible at this locality.

The most easterly exposure of the NCSM in the type area is in a railway cutting (ST36402820) 0.6 km east of Churley Farm (Fig. 1); there the unit comprises 2.5 m of greenish grey mudstones with subordinate greenish grey, fine-grained thinly bedded sandstones. The member is traceable by feature for only about 1 km farther eastwards, towards Stathe.

Along the strike section afforded by its outcrop in the type area, the NCSM shows considerable variation in thickness and lithology. The thinner developments in the eastern part of the type area and to the west of Knapp are composed largely of mudstones with siltstones and thin or impersistent sandstone beds. In the intervening area, between Mare Green and Knapp, the sequence is thicker and characterised by relatively substantial beds of sandstone, as exemplified in the type section (Table 1).

Palaeontology

Trace fossils

Ussher (1908, p. 59) observed "sandy tube-like markings which may be of organic origin" in mudstones in the NCSM near Moredon. These markings comprise small sub-horizontal sand-filled tubular burrows; they are locally abundant and are commonest in mudstones interbedded with thin siltstones and fine sandstones. The burrows represent an infaunal feeding trace of *Phycodes* or *Planolites* type (J. E. Pollard, *pers. comm.*) and are currently being studied by Dr. J. E. Pollard at Manchester University as part of a review of the British Triassic ichnofauna.

Macro fossils

Moore (1861, p. 486) recorded "*Estheria*, Plants, traces of fish scales, and a Reptilian bone" from exposures of the NCSM near North Curry (c. ST319256); "*Estheria minuta*" is recorded from mudstones in the member near Moredon House, North Curry (Ussher 1908, p. 59). Farther west, at Ruishton (c. ST265245), the member yielded "teeth of *Labyrinthodon*, serrated teeth of *Belodon*, *Acrodus Keuperinus*, &c., with *Estheria minuta*" (Moore 1867, p. 468).

Subsequently (Moore 1880, p. 198), the components of this fauna were listed as "teeth of *Thecodontosaurus* and *Palaeosaurus, Batrachia, Sphenonchus* and *Diplodus* ... and ... a form which might be named *Triplodus*", and (*ibid* 1881, p, 81) "teeth of *Thecodontosaurus* ... also *Acrodus Keuperinus, Hybodus, Diplodus*, &c.". The teeth assigned by Moore to *Diplodus,* and described, as *D. moorei*, by Woodward (1889, pp 299-300),'have been transferred to the genus *Xenacanthus* by Johnson (1980).

Palynomorphs

Organic residues from 23 samples from five sections in the NCSM and contiguous beds in the Mercia Mudstone Group to the east of North Curry have been examined for palynomorphs.

The NCSM was sampled at the following sites:

1. Type section; lane leading to Moredon House MPA 15874; above prominent sandstone unit (ST32542587) MPA 15875; below prominent sandstone unit (ST32552580)

2. South side of lane near Dyke's Farm MPA 15886 (ST35112769) MPA 15887 (ST35132773)

3. West side of Gainsload Drove (ST35692814) Section sampled from base (MPA 15876) upwards (to MPA 15881) at 0.5 m intervals.

4. Railway cutting (ST364282)
   SAL 1816; north-east side of cutting
   SAL 1817, 1818; south-west side of cutting

Beds in the Mercia Mudstone Group overlying the NCSM were sampled successively between ST35702812 (MPA 15885) and ST35702811 (MPA 15882) in Gainsload Drove.

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Samples from the Mercia Mudstone Group below the NCSM were mostly devoid of palynomorphs other than modern contaminants or, in the case of MPA 15871, lycospores and densospores reworked from Carboniferous deposits. A few poorly-preserved and indeterminate but possibly indigenous (i.e. Triassic) bisaccate miospores were, however, recovered from the lowest of these samples (MPA 15868).

Samples from the NCSM and overlying beds were, with the exceptions of MPA 15886, SAL 1817 and SAL 1818, productive and, with the exception of the highest sample from beds above the NCSM (MPA 15882), yielded fairly diverse and generally profuse but poorly preserved assemblages of miospores associated with fragmentary remains of land plants (e.g. cuticles, vascular tissues, etc.). Remains of *Plaesiodictyon mosellanum* Wille 1970, a possible green alga of the order *Chlorococcales*, were observed in preparations from Gainsload Drove (MPA 15876, 15877, 15883) and the nearby railway cutting (SAL 1816).

The miospore associations (Fig. 2) are dominated by bisaccate, principally disacciatrilete, taxa which, though commonly indeterminate because of poor preservation, include representatives of the genera *Alisporites* and, in smaller numbers, *Cuneatisporites, Platysaccus, Rimaespontes* and *Vesicaspora*. Other groups of miospores are slightly better represented in preparations from the NCSM of the type section and at Gainsload Drove than in those from the remaining localities and some of the overlying beds at Gainsload Drove.

Noteworthy features of the assemblages are the consistent presence of *Alisporites* cf. *parvus* de Jersey 1962, *Camerosporites secatus* Leschik *emend*. Scheufing 1978, Duplicisponites granulatus Leschik emend. Scheuring 1970, D. verrucosus Leschik emend. Scheuring 1978, *Ellipsovelatisporites plicatus* Klaus 1960, *Ovalipollis pseudoalatus* (Thiergart) Schuurman 1976 and *Vallasporites ignacii* Leschik 1955 *sensu* Scheuring 1970, and the comparative scarcity and sporadic occurrence of taxa of trilete miospores (e.g. *Verrucosisporites*) and disaccitriletes (*Triadispora*).

Taeniate bisaccates (*Lunatisporites*) and representatives of the genera *Aratrisporites, Patinasporites* and *Brodispora* were recorded sporadically from the NGSM and overlying beds in the Gainsload Drove section (Fig. 2).

Dating and correlation

The miospore assemblages from the NGSM and immediately overlying beds are, by comparison with those from other European successions from which independent biostratigraphic evidence is available, of late

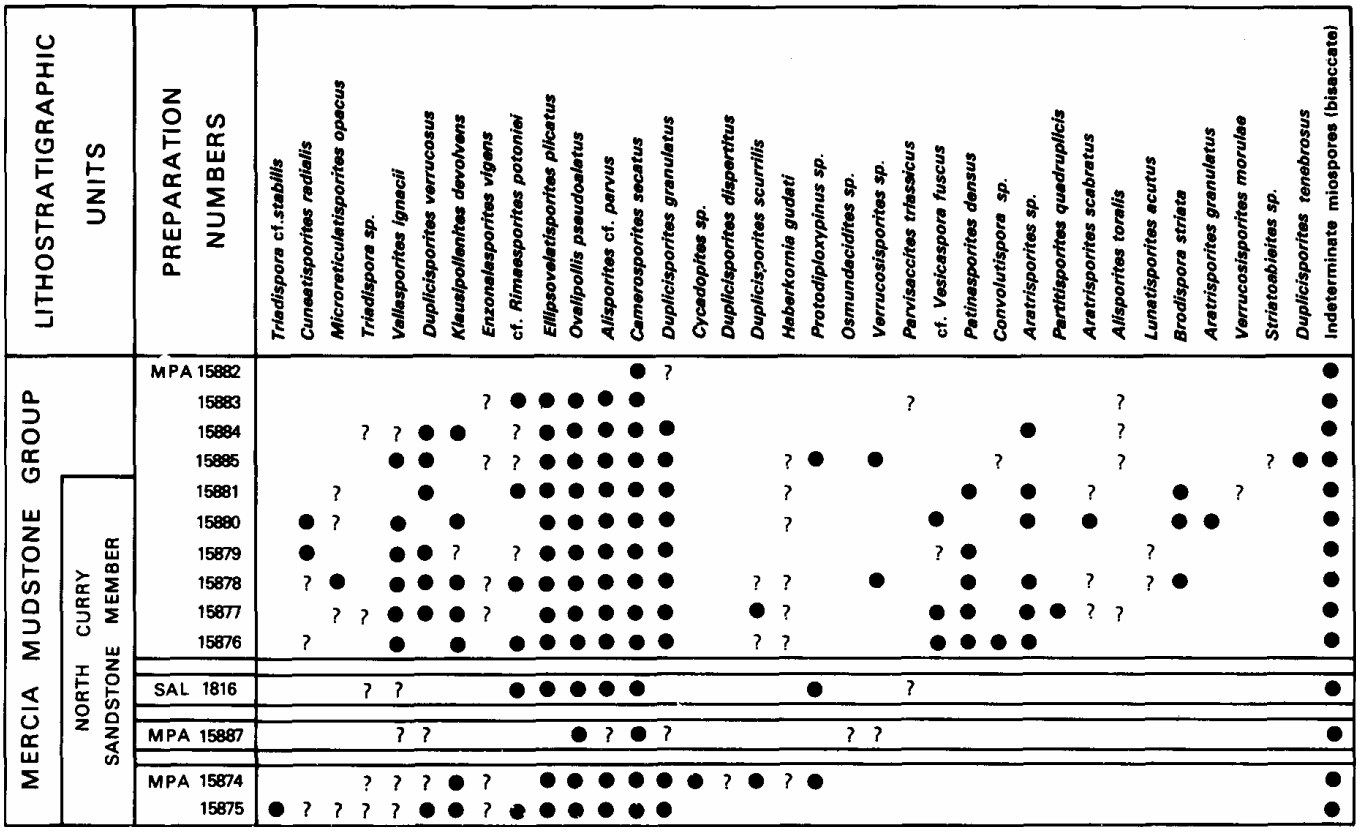

Figure 2. Miospores recorded from the North Curry Sandstone Member, and contiguous (overlying) beds in the Mercia Mudstone Group, in the type area.
Preparations and slides are held in the micropalaeontological collections at the British Geological Survey, Keyworth, Notts., and are registered in the MPA series.

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Triassic, Carnian, age. Comparison with results from Ladinian and Carnian sequences in the Dolomites and Swiss Alps (Scheuring 1970, 1978; Mostler and Scheuring 1974; van der Eem 1983) and Sicily (Visscher and Krystyn 1978) indicates that the North Curry assemblages, which contain *Vallasporites ignacii*, *Patinasporites densus* and *Duplicisonites* spp. in association with *Camerospontes secatus* and *Ovalipollis pseudoalatus*, but which lack *Echinitosporites iliacoides* Schulz and Krutzsch 1961, are no older than early Carnian (Cordevolian) and are likely to be later Carnian (i.e. Julian or Tuvalian) in age. The assemblages are comparable in composition with those known from the Arden Sandstone and Weston Mouth Sandstone members in the Mercia Mudstone Group successions of, respectively, Worcestershire and Devon (Clarke 1965; Warrington 1970, 1971; Fisher 1972, in Jeans 1978) and are indicative of a correlation with those units.

The macrofossils recorded from the NCSM (Moore *op. cit.*; Winwood 1906; Ussher *op. cit.*; Duffin 1978; Johnson 1980) are either unsuitable for precise age determination or require re-examination of the original material to be useful for this purpose. The assemblage, comprising plant remains, trace fossils, branchiopod crustaceans and the remains of fish, amphibians and reptiles, is, however, comparable with those known from the analogous late Triassic Arden Sandstone, Dane Hills Sandstone and Weston Mouth Sandstone members present at similar levels in the Mercia Mudstone Group successions of central and southern England (Warrington 1976; Warrington *et al.* 1980) and supports the correlation with those units which is indicated by palynological work. The possibility of correlation of the NCSM with the Arden Sandstone Member was recognised by Moore (1880, p. 198), and Lott and others (1982), in a study of the Mercia Mudstone Group in the western Wessex Basin, regarded the NCSM as a correlative of the Weston Mouth Sandstone Member.

Conditions of deposition

Scarcity of exposures hampers the study and interpretation of the NCSM. However, thick cross-bedded sandstones are present in the central part of the outcrop around North Curry. and Knapp and pass laterally, to both east and west, into thinner sequences of interbedded mudstones, siltstones and fine sandstones. Interpretation of the overall geometry of these deposits is impeded by the limited (strike section) nature of the outcrop but the thick sandstones probably occupy the site of a distributary or channel that was flanked by overbank or mudflat areas in which the thinner, more argillaceous, laterally equivalent sequences, characterised by abundant bioturbation, accumulated. No satisfactory palaeocurrent vector measurements have been obtained from these deposits, but a south-easterly direction of transport is inferred from cross-bedding observed by one of us (BJ.W.) in sandstones in the type section near Moredon House.

The fossils recorded from NCSM include forms indicative of both terrestrial and aquatic conditions. The former include miospores and indeterminate remains of components of the parent land flora, and teeth of reptilian origin ("*Palaeosaurus*" and "*Thecodontosaurus*"). Organisms inhabiting or associated with an aqueous environment comprise possible green algae (*Plaesiodictyon*; Wille 1970), branchiopod crustaceans (*Euestheria*) and fish, the latter represented largely by dermal and dental remains including *Sphenonchus* (cephalic spines of male hybodont sharks; Duffin 1978) and teeth of hybodont and xenacanthiform sharks (*Acrodus*, *Hybodus* and *Xenacanthus*). The presence of amphibians is indicated by a record (Moore 1867, p. 468) of "*Labyrinthodon*", and an aquatic reptile is represented by teeth identified as "*Belodon*?", a phytosaur (Duffin 1978). The water inhabited by these organisms may, from the presence of *Plaesiodictyon*, *Euestheria*, *Xenacanthus*, an amphibian and a phytosaur, have been largely brackish with connection to a marine source.

The NCSM appears, from the limited data available, to have formed in a deltaic or estuarine environment comparable with that envisaged for the Schilfsandstein, a unit of similar facies and Carnian age in the Triassic succession in Germany (Wurster 1964a b); probable equivalence with the "*Schilf*" sandstone of the Continent was alluded to by Ussher (1906, p.22).

Acknowledgements. The writers are grateful to J. E. Pollard (Manchester University) for his comments on trace fossils observed in the NCSM, and to R. W. Gallois, H. C. Ivimey-Cook, M. Mitchell and A. Whittaker (British Geological Survey), G. W. Green (Nottingham) and C. J. Duffin (Morden) for their comments upon an initial draft of this account. F. S. Fisher, of Drayton, is thanked by one of us (G. W.) for assistance in the field.

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Read at the Annual Conference of the Ussher Society, January 1984

Possible wave-influenced sedimentary structures in the Bude Formation (Lower Westphalian, south-west England), and their environmental implications

R. HIGGS

Higgs, R. 1984. Possible wave-influenced sedimentary structures in the Bude Formation (Lower Westphalian, south-west England), and their environmental implications. *Proceedings of the Ussher Society, 6, 88-94.*

The Bude Formation, whose depositional environment has long been uncertain, is considered to be deltaic by some authors, and of deep-water-fan origin by others. This paper documents an apparent abundance of wave-influenced sedimentary structures in the Bude Formation. Many sharp-based, graded beds are capped by ripples and ripple cross-lamination showing characteristics suggestive of wave-action. A structure thought to be hummocky cross-stratification, which is now widely considered to reflect storm-wave-induced oscillatory flow, underlies the ripple cross-lamination in some of these beds, but only rarely is it clearly expressed. It is suggested that much of the Bude Formation was deposited in relatively shallow water, between storm wave base and fair-weather wave base.

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Introduction

The Lower Westphalian Bude Formation, partially equivalent to the deltaic Bideford Formation of north Devon (Edmonds et al. 1979), but lacking the obvious cyclicity of the latter, is thought to be about 1300 m thick (Freshney and Taylor 1972), and crops out over approximately 1000 km2 in north Devon and north Cornwall (Fig. 1). Late Carboniferous (Variscan) orogenic compression crumpled the strata into a complex of mainly upright, open folds, with sub-horizontal, east-west trending axes, and wavelengths ranging from tens of metres to kilometres (Freshney et al. 1979). The Formation conformably overlies several hundred metres of shales with relatively thin, turbidite-like sandstones (the Crackington Formation; ibid.), and is unconformably overlain by the post-orogenic New Red Sandstone (mainly Permo-Triassic). Although inland exposure is poor, the north-south stretch of coastline between Widemouth Sand and Hartland Point (Fig. 1) provides magnificent exposures in near-continuous cliffs and wave-cut platforms, and affords an opportunity to examine the entire Formation in section perpendicular to strike.

A detailed facies analysis of the Bude Formation is currently being conducted by the author. There appear to be two main facies, interbedded with one another. The first of these facies comprises sharp-based, graded, very fine sandstones; these are evidently the product of discrete, waning-energy depositional events, and may therefore be termed "event deposits" (Seilacher 1982). Individual event deposits are generally less than 30 cm thick, but they are commonly amalgamated (mainly by thorough syn-depositional loading of intervening muds), forming sandstone bodies up to 10 m thick. The other main facies consists of mudstones and shales; these occur in layers which are mostly less than lm thick, with the exception of five or six nodular shales up to 10 m thick (Freshney et al. 1979). The Bude Formation shows no obvious thickening-upward or thinning-upward sequences, although this may be an artefact of the unusual style of amalgamation (see above). Most of the event deposits show a preferred sequence of internal structures, as follows: sharp base + massive texture +- horizontal lamination + asymmetrical ripple cross-lamination. An abrupt gradation from very fine sandstone to siltstone is usually visible close to the top of the bed; nevertheless, the contact with overlying mudstone or shale is normally sharp. Palaeocurrent analysis using unidirectional sole marks indicates that the sediments were derived from all quadrants except the south (ibid.). Body fossils are very scarce, and are totally lacking in the event deposits, even as moulds. Goniatites and pelagic bivalves are the only definite marine fossils found in the Bude Formation; they occur in discrete bands within a few of the thick nodular shales (ibid.). Other body fossils are limited to fish and a single crustacean (ibid,; White 1939); these, too, are restricted to the nodular shales. Of more widespread stratigraphic occurrence, but still mainly confined to mudstones and shales, are trace fossils, including the walking tracks of xiphosurids (i.e. horseshoe crabs; Goldring and Seilacher, 1971), and burrows, particularly *Planolites* (King, 1966) and *Diplocraterion*.

There has been a long-standing controversy over the depositional environment of the Bude Formation (Table 1; Higgs 1983). Owen (1934) considered an abundance of plant detritus and the presence of "wash-out channels" (ibid.) to indicate a deltaic environment. King (1967) also diagnosed deltaic conditions, citing the presence of xiphosurid tracks as evidence for "shallow water". In contrast, Ashwin (1957) postulated deposition at the foot of a continental slope, pointing out that many of the thinner sandstones exhibit features typical of turbidites, including grading, sole marks, and a characteristic sequence of internal structures (see above). Burne (1969) and Melvin (1977) concurred that most of the sandstones

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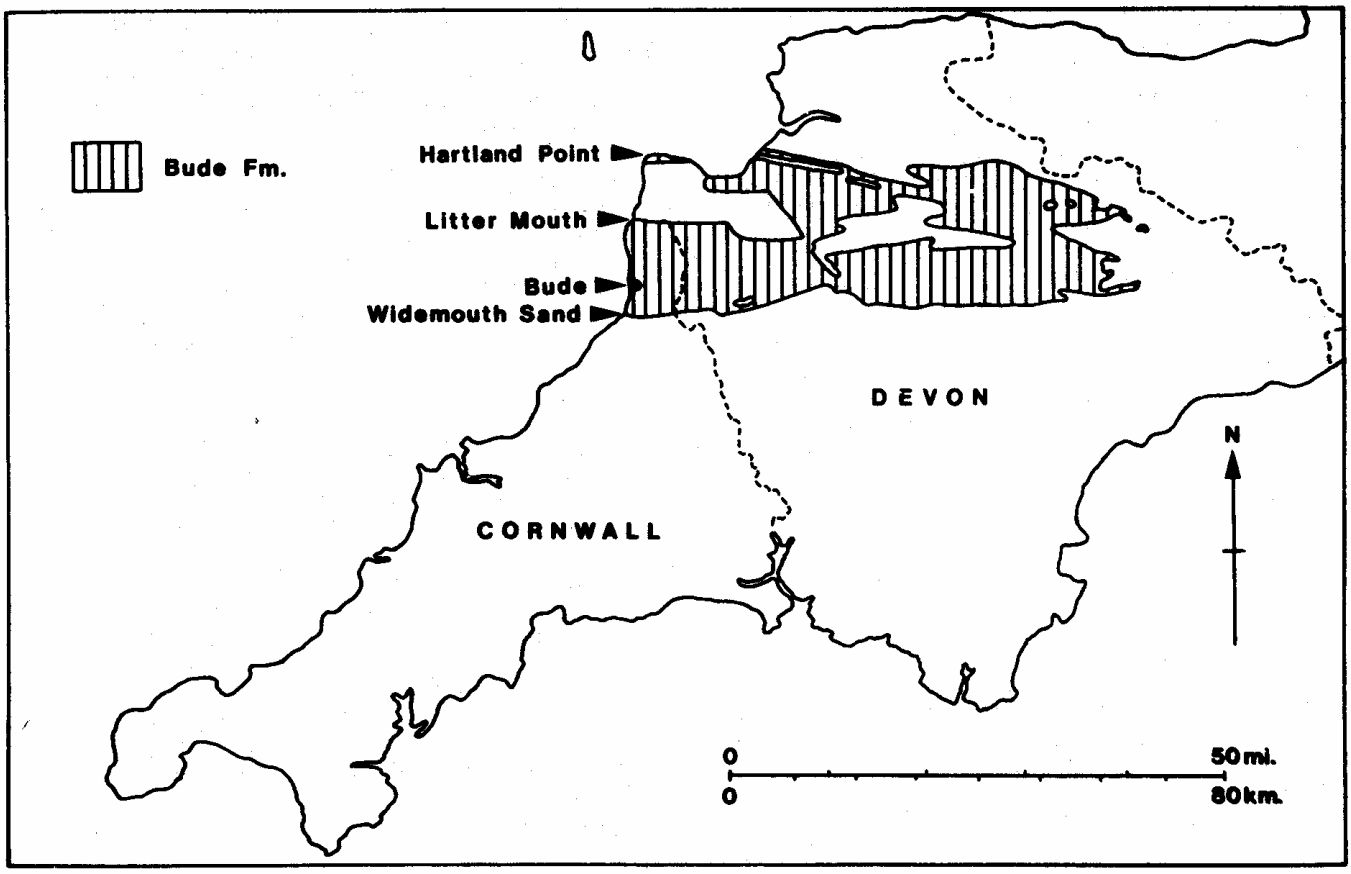

Figure 1. Location map, showing the outcrop of the Bude Formation.

AUTHOR	YEAR	ENVIRONMENT	REMARKS
Owen	1934	Delta	
Ashwin	1957	Foot of continental slope	First to suggest turbidity currents
Prentice	1962	Delta	
Lovell	1964	Continental slope	"greater part ... deposited by turbidity currents"
King	1967	Fluvial, deltaic, shelf	
Edmonds and others	1968	Delta	"Turbidites are rare or absent"
Burne	1969	"Relatively deep water" fan	"out of reach of wave activity"; c.f. La Jolla Fan
Goldring and Seilacher	1971	Lake floor or "outer slopes"	Lake "large, though probably not very deep"
Freshney and others	1972	Delta	Max. water depth about 60 m
Melvin	1977	"Relatively deep water" fan	Not "shallow, agitated water ... such as near-shore shelf, beach or delta"; not abyssal
Freshney and others	1979	Delta slope	Basin possibly "too narrow ... for deep water sedimentation"; depth only "a few metres" at times

Table 1. Chronological catalogue of previous interpretations of the depositional environment of the Bude Formation

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are turbidites, and both suggested deposition on a subaqueous fan in "relatively deep water".

Despite the "fundamental difference of opinion as to the processes of sedimentation" (Reading 1963), previous authors agree almost unanimously that evidence for wave-induced oscillatory flow is lacking. Excepting King (1967, p. 95), who listed "oscillation" ripples among the features from which he determined palaeocurrents (King failed to incorporate these ripples in his environmental interpretation), and Higgs (1983), who tentatively identified hummocky cross-stratification, there are apparently no previous reports of wave-influenced sedimentary structures in the Bude Formation.

However, in the light of significant advances, since the mid-1960s, in our ability to recognise wave-influenced structures (see, for example, review in Reineck and Singh 1980), it now appears that such structures are common in the Bude Formation. Having completed a sedimentological reconnaissance of the entire coastal section between Widemouth Sand and Litter Mouth (Fig. 1), encompassing all of the known stratigraphic range of the Formation (Freshney and Taylor 1972), the author is of the opinion that many of the ripples and much of the ripple cross-lamination occurring at the tops of event deposits throughout the succession are asymmetrical, wave-influenced varieties. In addition, a structure thought to be hummocky cross stratification, which is widely regarded as a product of intense oscillatory flow (Dott and Bourgeois 1982), was observed in a few beds. The characteristics and origins of each of these structures, and their environmental implications, are discussed below.

Ripples

Description

Well-exposed rippled bedding surfaces, representing the tops of event deposits, are a common feature of coastal exposures of the Bude Formation. Some of these surfaces show ripples with strongly asymmetrical, angular crest profiles, and sinuous to irregular crestlines. Most ripples, however, have straight to slightly sinuous crestlines, a slightly asymmetrical crest profile (ripple symmetry index generally 2 to 3), a uniform spacing of between 5 and 15 cm, and a uniform height of 1 to 3 cm (ripple index generally 5 to 8; see Reineck and Singh (1980) for parameter definitions); crest profiles are usually rounded, and crestlines are continuous, with rare bifurcations (Fig. 2).

Interpretation

Previous workers, without exception, have evidently interpreted both types of ripple described above as "current ripples", which are those produced by steady unidirectional flows (Harms et al. 1982). However, while agreeing with this interpretation for the first type, the author submits that the ripples of the second type are intermediate in character between typical strongly asymmetrical, sinuous- to irregular-crested current ripples, and the symmetrical, straight-crested "wave ripples" produced by those oscillatory flows which are themselves "symmetrical" in that no net sediment translation occurs (Harms 1979). Such bed-forms have been referred to as "intermediate flow ripples" by Harms et al. (1982, p. 2-45): these can form either (1) when waves are "rapidly shoaling" (ibid., p. 2-3), whereupon the resulting oscillatory motions become strongly asymmetrical, causing net unidirectional sediment transport; or (2) when waves operate in the presence of an independent unidirectional current (resulting in a "combined flow"; ibid.). There is a strong resemblance to the "wave-dominated combined-flow ripples" of Harms (1969); However, it is important to note that Harms' experimental investigations used medium sand, whereas most of the Bude Formation ripples occur in siltstone. Unfortunately, there appears to be a complete lack of data, both from the laboratory and from natural environments, on combined-flow ripples in silts (Harms et al. 1982). Nevertheless, the author feels that, by analogy with sands, there is considerable justification for referring to the ripples in question as "wave-influenced" forms.

Ripple cross-lamination

Description

Ripple cross-lamination is common throughout the Bude Formation; it occurs primarily in silt-grade sediment, both in the uppermost few centimetres of relatively thick event deposits, and throughout the entire thickness of relatively thin ones. Walker (1963) recorded a variety of "ripple-drift cross-lamination", in which any one lamina, followed laterally, traces a train of asymmetrical ripples representing a former depositional surface. However, there is a more common type of cross-lamination which, when viewed perpendicular to ripple crestlines, appears as a pile of interleaved, lenticular to irregular sets, with scoop-shaped to undulating, erosional set boundaries, and with a predominantly unidirectional internal lamination (Fig. 3). Sets commonly appear to climb, one upon another, at a low angle. Sets are usually a few centimetres to a few decimetres long, and up to 2 cm thick. When traced downward within a set, the cross-laminae decrease in angle of dip, from a maximum of about 45°, and meet the lower set boundary tangentially; in many cases, descending laminae actually swing through the horizontal, and then ascend the flank of the adjacent set as "offshoots" (Fig. 3; Raaf *et al.* 1977).

Interpretation

Previous authors have invariably implied that all of the ripple cross-lamination in the Bude Formation can be attributed to unidirectional currents. However, while such an origin may be valid for the first of the two varieties of cross-lamination described above (Walker 1963), the author contends that it is probably inapplicable to the second, more common, type. The non-tabular sets, strongly curved set boundaries, and offshooting laminae typical of the latter are all suggestive of wave-induced oscillatory flow, as has been demonstrated by Boersma (in Raaf *et al.* 1977). Hence, it appears reasonable to refer to this structure as "wave-influenced ripple cross-lamination". If this interpretation is correct, the overall unidirectionality of the cross-lamination, reflecting net one-way transport of sediment, would indicate either that the waves were "rapidly shoaling" (Harms et al. 1982), in which case the resulting oscillatory flow would have been strongly asymmetrical

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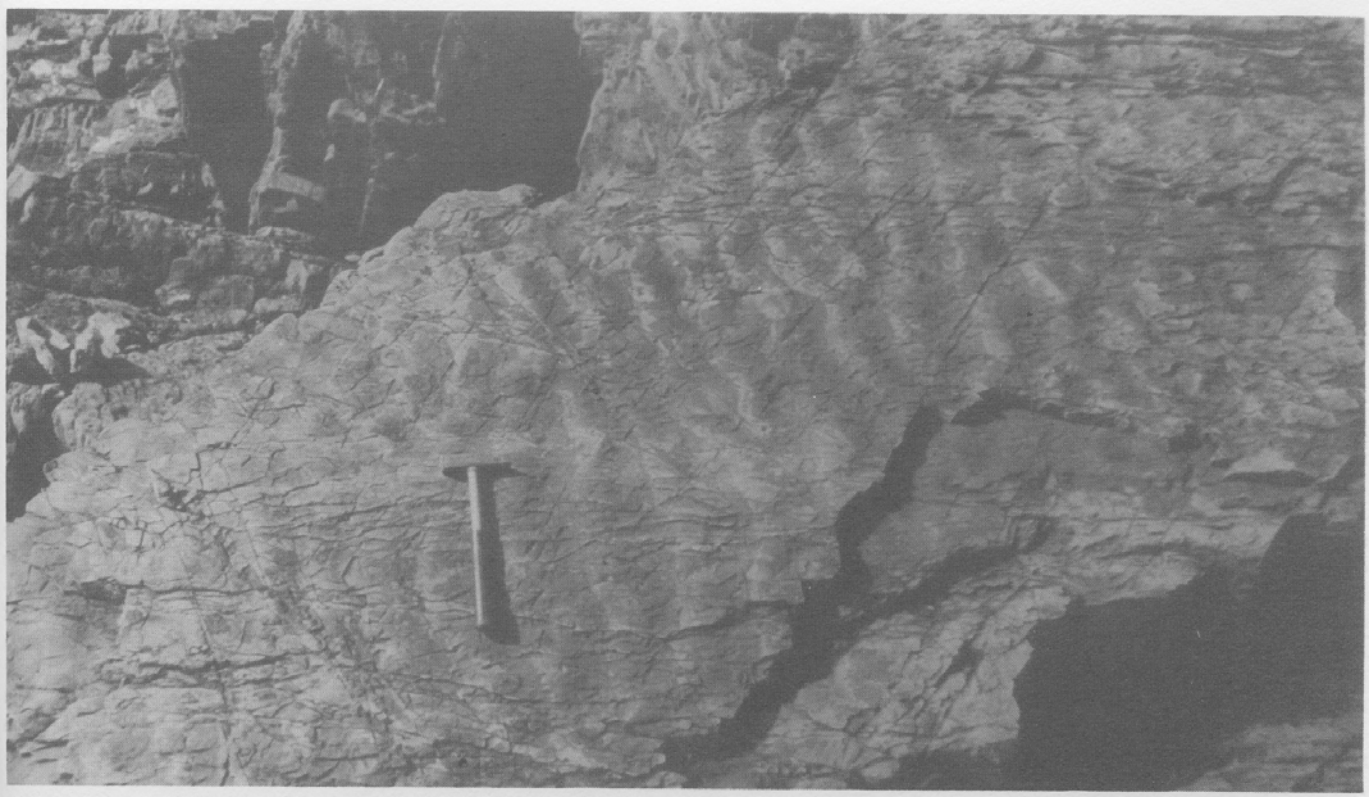

Figure 2. Upper surface of a Bude Formation event deposit, showing probable wave-influenced ripples. Note slightly sinuous crestlines, roughly symmetrical crest profiles, and crestline bifurcation (immediately above and right of centre). Grain size is silt. Hammer 30 cm long. Cliff base, SS 20151048.

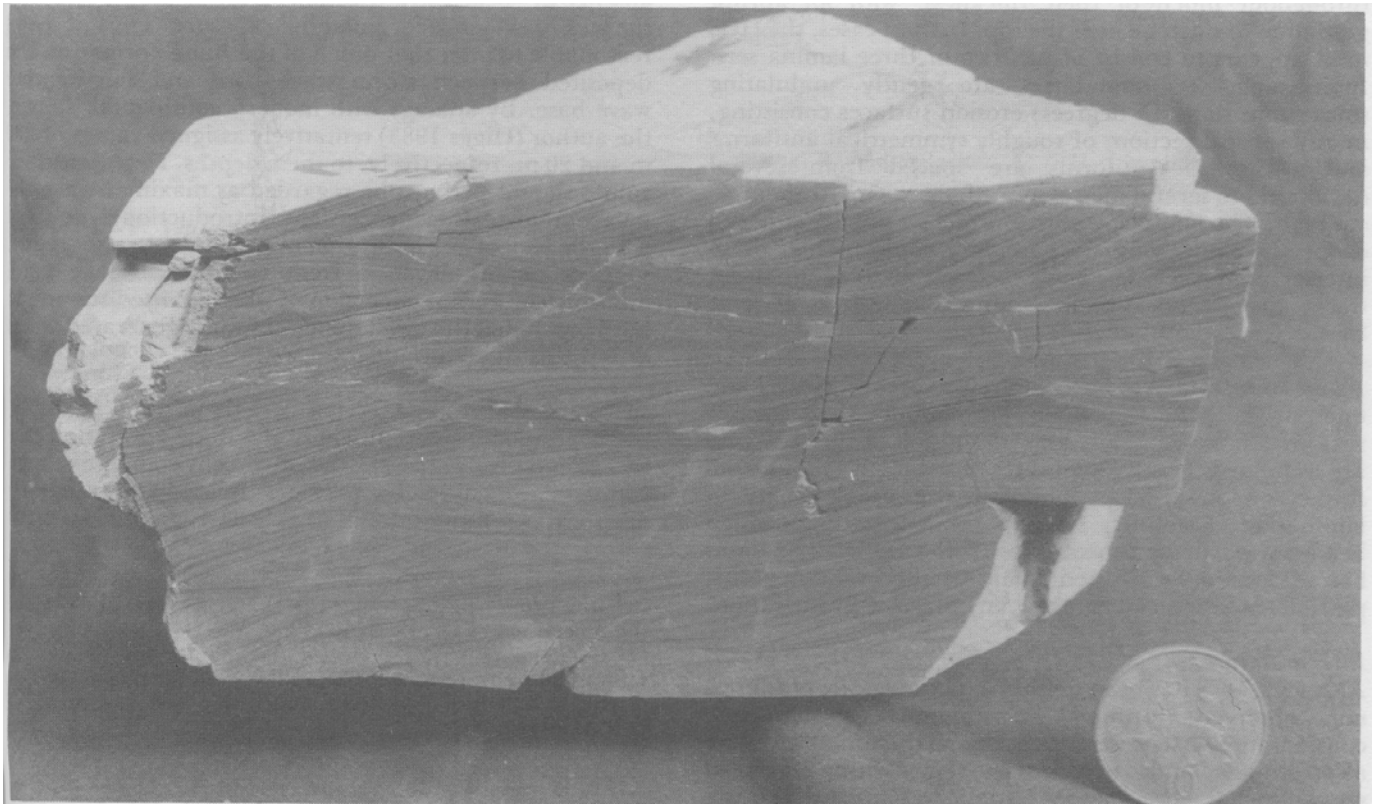

Figure 3. Bude Formation polished sample showing probable wave-influenced ripple cross-lamination. Sample shows the entire thickness of a single event deposit. An abrupt gradation from very fine sandstone to siltstone is discernible about 1 cm above the base. Note lenticular sets, strongly curved set boundaries, and "offshoots" (see text). Sample right way up. Polished face almost perpendicular to ripple crestlines. Sample taken from cliff base at SS 19990494.

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(ibid.), or that the waves were superimposed upon an independent unidirectional current. (Not surprisingly, both of these alternatives also seem to be plausible with regard to the origin of the dominant ripple type in the Bude Formation; see previous section). The fact that set boundaries are erosional indicates that ripple migration was rapid, relative to the rate of sedimentation, enabling scour within the ripple troughs to erode sediment previously deposited on the flanks of adjacent crests.

Hummocky cross-stratification

Description

Hummocky cross-stratification (HCS), first defined by Harms et al. (1975), was tentatively identified in the Bude Formation by the author (Higgs 1983); it is felt that subsequent field observations have confirmed its presence (Fig. 4). Unequivocal examples of the structure are rare, having been observed in only about ten individual event deposits to date. However, there are many beds in which a structure which may originally have been HCS has been obscured by soft-sediment deformation. Beds with HCS apparently occur unpredictably throughout the Bude Formation, both singly, and within amalgamated-sandstone units. Individual HCS-bearing event deposits are about 20 to 40 cm thick, and may show a massive interval at the base, and a few centimetres of wave-influenced ripple cross-lamination (see previous section) at the top. Most of the beds consist of very fine sandstone throughout much of their thickness, with an abrupt gradation to siltstone near the top. In most cases, the HCS itself appears to consist of just two or three lamina sets. Intervening set boundaries are gently undulating (maximum slope 15 degrees) erosion surfaces consisting, in any vertical section, of roughly symmetrical antiforms and synforms. Antiforms are spaced from several decimetres to several metres apart, and are up to about 30 cm high. The laminae within any one set appear either to drape the underlying set boundary conformably, or to intersect it at a very low angle. Sets are up to about 30 cm thick, and extend laterally from a few decimetres to at least a few metres.

As mentioned above HCS is rarely expressed clearly and unequivocally in the Bude Formation. Two possible reasons for the generally poor definition of the structure are as follows. Firstly, high rates of sediment fallout from suspension may have inhibited the development of lamination Secondly, syn-depositional loading, which has been particularly extensive in the Bude Formation, has undoubtedly affected many of the event deposits, partially or completely erasing any primary stratification.

Interpretation

Although HCS has been neither experimentally produced nor observed forming m modern environments, consideration of its stratification characteristics, facies associations, and overall position within vertical sequences has led numerous authors to suggest that it is a product of intense, storm-wave-generated, oscillatory flow (Harms et al. 1975, 1982; Dott and Bourgeois 1982). However, it is not generally appreciated that during a storm, wave height may vary considerably from one wave to the next (Draper 1967). This fact raises the possibility that, during the emplacement of a hummocky cross-stratified bed, it is only the erosional set boundaries which are formed by intense oscillatory flow, during the passage of a particularly large wave or group of large waves. Deposition of the intervening sets of laminae, on the other hand, may take place while the passing waves are relatively low, when the only effective current is some kind of sediment-supplying unidirectional flow unrelated to the waves.

Environmental implications

Wave-influenced ripples and corresponding cross-lamination appear to be common at the tops of event deposits throughout the Bude Formation, suggesting that the beds concerned were deposited above storm wave base, where wave-induced oscillatory flow was able to influence bottom-sediment movement. Hummocky cross-stratification also seems to be present in a number of event deposits. This structure is widely considered to form between storm wave base and the shoreline under the influence of intense oscillatory flow produced by storm waves (Harms et al. 1975, 1982; Dott and Bourgeois 1982). However, according to Dott and Bourgeois (1982), preservation is unlikely above fair-weather wave base. Deposition of the Bude Formation beneath fair-weather wave base is suggested by the absence of trough- and tabular cross-stratification, and is consistent with the absence of "beach stratification" (Harms 1979), and with the lack of evidence for subaerial exposure. Thus, it seems reasonable to infer that much of the Bude Formation was deposited between storm wave base and fair-weather wave base. By analogy with modern continental shelves, the author (Higgs 1983) tentatively assigned values of 200 m and 20 m, respectively, to these depths. However, these values should perhaps be regarded as maxima, since the palaeontological evidence (see Introduction), coupled with the lack of any indication of proximity to a shoreline, suggests deposition in a fresh to brackish lake (with intermittent marine incursions), whose limited size might have precluded the development of wind-generated waves as large as those in modern oceans.

In conclusion, it appears that, strictly in terms of depth, the depositional environment of the Bude Formation may fit essentially between the two environments most commonly invoked by previous investigators, namely deep-water fans and deltas (Table 1; Higgs 1983).

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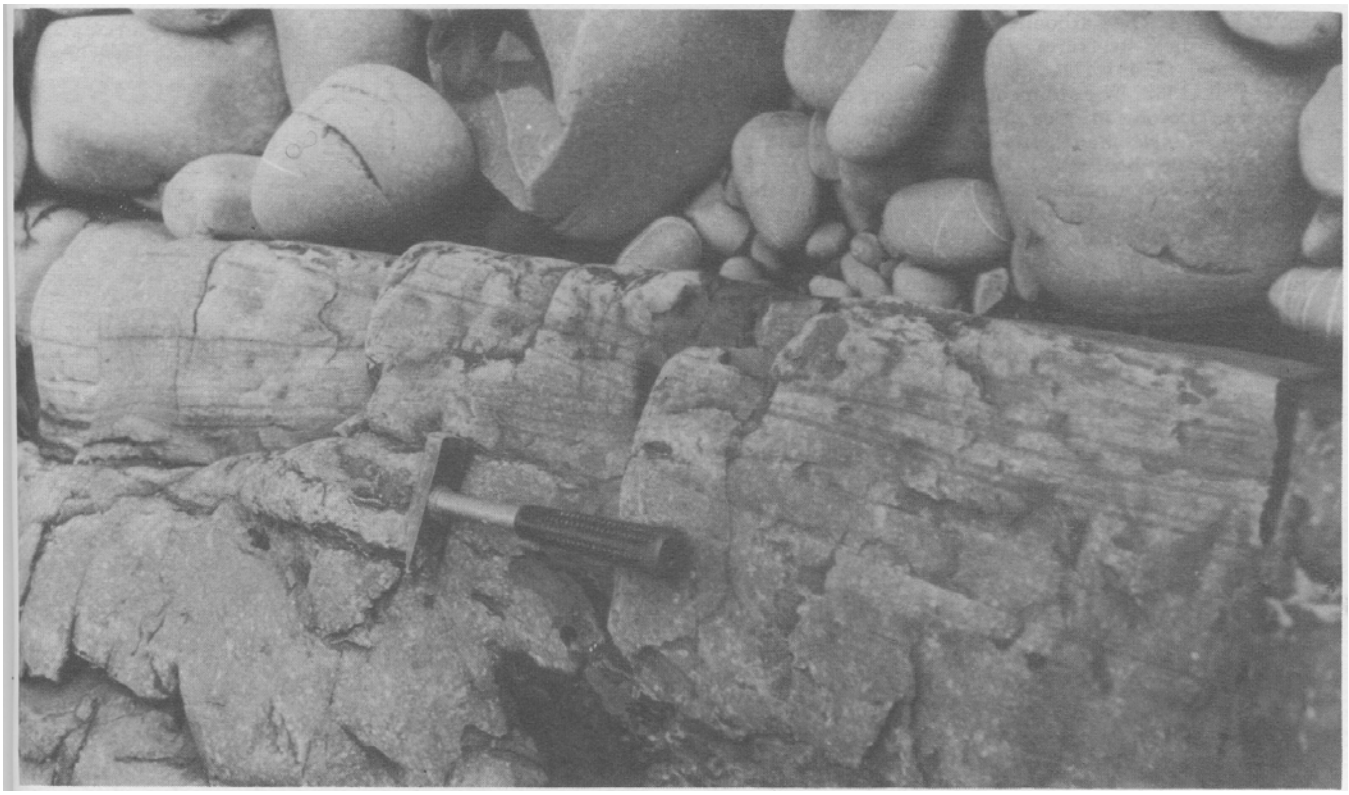

Figure 4. Probable hummocky cross-stratification in very fine sandstone, Bude Formation. Note: (1) set boundaries are low-angle erosion surfaces, conformably draped by lamination which appears to thicken laterally in places; (2) upper surface of sandstone unit appears erosional, and shows slight upward convexity, suggesting possible hummocky topography; and (3) sandstone in bottom one-third of photograph appears largely structureless (see text). Section youngs towards top of photograph. Hammer 30 cm long. Wave-cut platform, SS 19940557.

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Read at the Annual Conference of the Ussher Society, January 1984

Diagenetic modifications of primary sedimentological fabric in the Westbury Formation (Upper Triassic) of St. Audrie's Bay, north Somerset.

J.H. S. MACQUAKER

Macquaker, J. H. S. 1984. Diagenetic modifications of primary sedimentological fabrics in the Westbury Formation (Upper Triassic) of St. Audrie's Bay, North Somerset. *Proceedings of the Ussher Society, 6*, 95-99.

A two stage model is proposed to account for the alternating rhythmic black shale/limestone sequence present in the Westbury Formation at St. Audrie's Bay. Stage 1 involves episodic storm sedimentation interrupting quiet water conditions, causing an influx of coarse lithoclastic debris and reworking bioclastic material. Stage 2 involves bacterially controlled diagenesis within the anoxic black shales mobilizing carbonate (Curtis 1980).

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Introduction

The nature of the Rhaetian/Hettangian transgression has interested geologists for many years. The section at St. Audrie's Bay has been chosen as part of a wider study to highlight aspects of the transgression and to test previous models discussing its form. In particular diagenetic fabrics were examined using a cathodo-luminoscope.

St. Audrie's Bay is located in north Somerset 5 km., east of Watchet (ST 103432) (Fig. 1). It is one of numerous Mesozoic exposures on the Severn Estuary. The section is exposed close to the western end of the St. Audrie's landslip.

Stratigraphy

The section exposes sediments of Norian to Hettangian age, including the Mercia Mudstone Group, the Penarth Group and the Lower Lias. Nomenclature follows the usage of Warrington et al. (1980).

Palaeoenvironmentally the sequence charts the change from predominantly terrestrial conditions, with hypersaline lake development (Jeans, 1978) in the Blue Anchor Formation, to predominantly marine conditions, in the Westbury Formation.

The Westbury Formation comprises an alternating sequence of carbonate rich horizons, including arenaceous lithoclastic and bioclastic debris and carbonate poor horizons, dominated by detrital silt and clay (Fig. 2).

Different models of environmental cyclicity have been invoked to explain this rhythmic alternating limestone and shale sequence. Models include, fluctuating energy conditions caused by sea level change, (Hallam, 1975; Whittaker and Green, 1983; and Hamilton 1977) and storm activity (Whittaker and Green, 1983; and Mayall, 1979).

Specific interpretations have applied to particular horizons, notably the "bone-bed", which has been interpreted as a "mass-mortality' event (see Antia, 1979 for review) and as a strand line deposit, (Hamilton, 1977; Whittaker and Green, 1983). Thicker sand dominated horizons have been ascribed to migrating barrier bar forms (Mayall, 1979).

Sample Collection and Analysis

Samples were collected from most of the major facies in the succession (Fig. 2). Whole rock and clay analyses were undertaken using a Philips X-ray diffractometer, with a

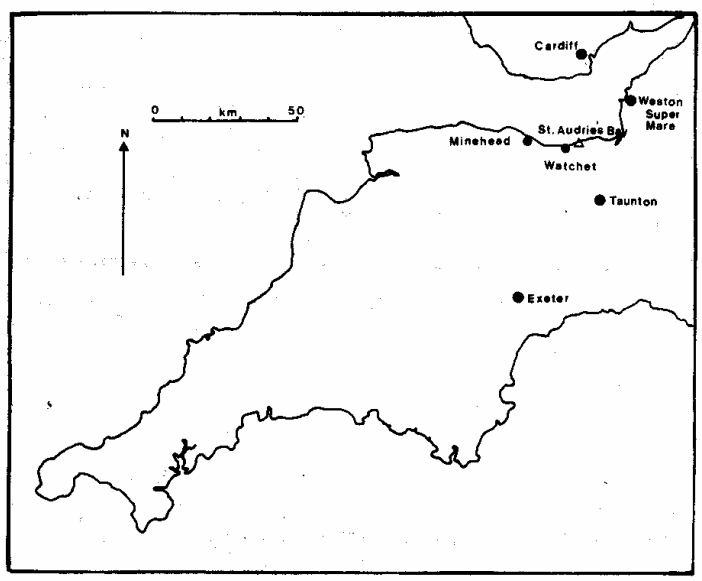

Figure 1. Location Map.

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variable slit, using copper K-alpha radiation, at 40 Kv. and 40 Ma. scanning at one degree two theta. Sample preparation was after Jeans (1978).

The cemented horizons were sectioned and stained with alizarin red S and potassium ferricyanide, after Dickson (1966). Subsequently they were polished using 3 micron diamond paste and examined on a Nuclide cathodo-luminoscope; at 18Kv. (henceforth abbreviated to c.1.). All photographic recording was on Ektachrome 64, via a Zeiss microscope and a Nikon FE camera, with exposure times varying between 2 and 10 minutes. For further details of C.I. operation see Nickel (1978).

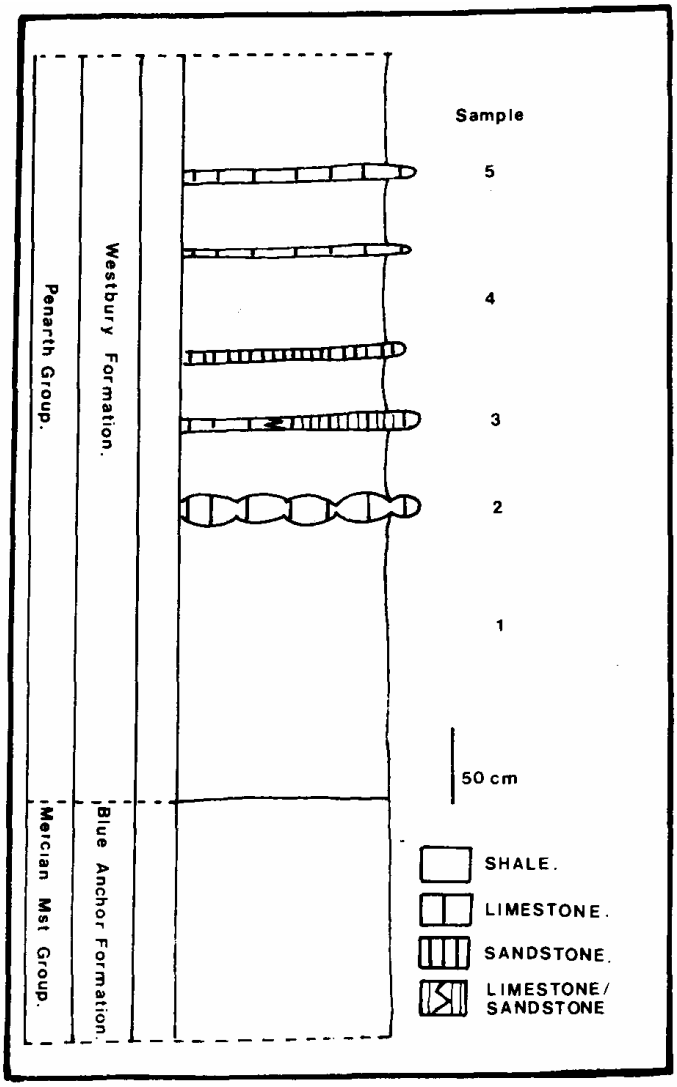

Figure 2. St Audrey's Bay, section and sample localities

Samples (Fig. 2)

St. Audrie's Bay 1 (STA 1).
Black unfossiliferous fissile mudstone, pyrite rich with silt grade detrital quartz and detrital clays, including illite and chlorite. No sedimentary structures are visible.

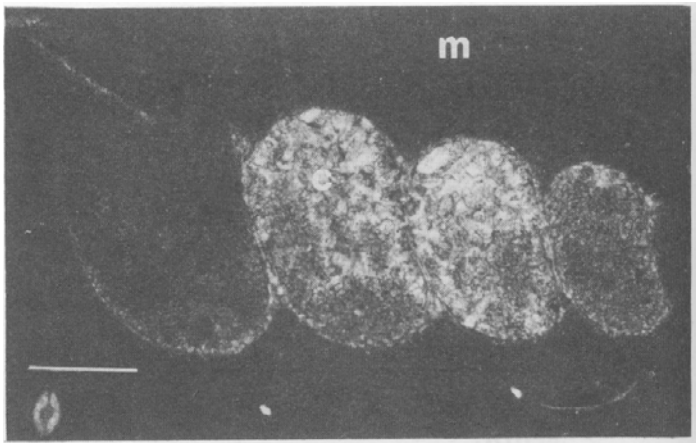

Figure 3. STA 2. Biomicrite with shelter porosity developed within gastropod whorls. m:micrite. Plane polarised light, Scale bar is 500 microns.

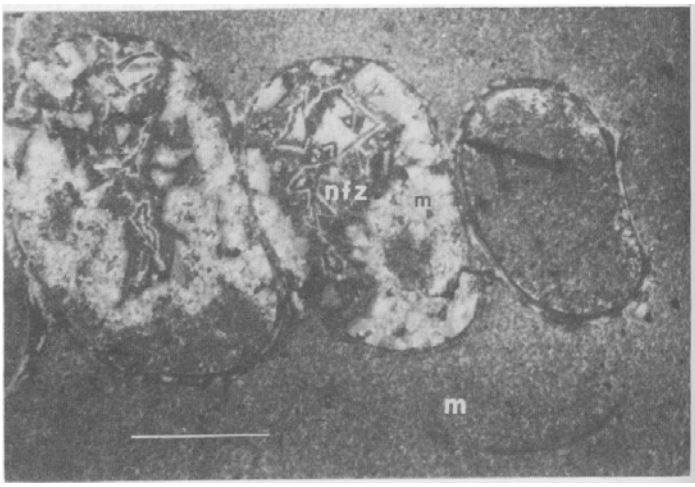

Figure 4. STA 2. (As figure 3) revealing zoned non-ferroan cement (nfz). c.l. conditions, scale bar is 500 microns.

St. Audrie`s Bay 2 (STA 2).
Calcareous mudstone. Nodular massive limestone, composed of micrite, with greater than 10% bioclasts, including recrystallised, reworked gastropods and pelecypods (*Protocardia rhaetica*).

The micrite is composed of non-ferroan calcite. Shelter porosity is developed within some of the bioclasts and has been infilled with multi-zoned c.1. non-ferroan cements (see Fig. 4) and ferroan cements (Fig. 3). Infilled joints show a similar pattern to the shelter porosity.

St. Audrie`s Bay 3 (STA 3)
Bioclastic packestone grading into a calcarenaceous sandstone. Grey thin pyritiferous limestone, with a symmetrically ripple marked top and a mud drape. Cone-in-cone is developed on the base. A poorly developed fining upward cycle is present with coarse vertebrate debris being concentrated near the bottom. Pelecypods are recrystallised (Fig. 5), disassociated and generally preserved convex up, without any other particular orientation. The fabric is aggraded micrite, preserved as non-ferroan calcite. Some shelter porosity is preserved, particularly beneath bioclasts (Fig. 5), containing c.1. zoned non-ferroan calcite and ferroan calcite (Fig. 6).

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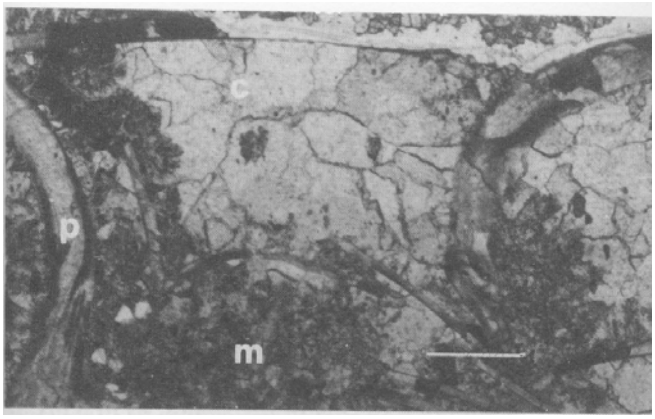

Figure 5. STAA 3. Bioclastic packestone with well developed shelter porosity infilled with ferroan (f) and non-ferroan (nf) cement. p:pelecypod; m:micrite. Plane polarised light, scale bar is 500 microns.

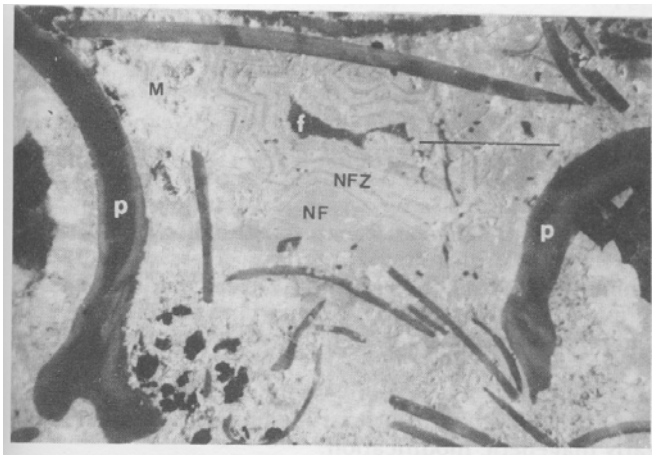

Figure 6. STA 3. Revealing both a non-zoned and a zoned cement phase and a non luminescing ferroan cement. c.l. conditions, scale bar is 500 microns.

St. Audrie's Bay 4 (STA 4)

Black fissile mudstone with alternating cemented horizons, pyrite rich and fossiliferous, with convex up and disassociated decalcified pelecypods including, *Protocardia rhaetica.* The horizon is dominated by detrital silt grade quartz and detrital clays, including illite and chlorite.

Figure 7. STA 5. Bioclastic grainstone with dominate pelecypod component. Plane polarised light, scale bar is 500 microns.

St. Audrie`s Bay 5 (STA 5)

Bioclastic grainstone. A grey fossiliferous limestone with pyrite, containing disassociated pelecypods, including *Modiolus sp.* and *Protocardia rhaetica*. The pelecypods are recrystallised and preserved as non-ferroan calcite, some are slightly micritized (Fig. 7). The cement is non-ferroan calcite.

Discussion

Primary sedimentological rhythms undoubtedly exist within the Westbury Formation; they are indicated by fine grained detrital fractions (low energy), being interspersed with coarser fractions (high energy), of bioclastic and lithoclastic debris. The fine grained component is characterized by having virtually no carbonate. Where pelecypods are present, they are invariably decalcified and only preserved as moulds. The fine grained horizons are pyrite rich and lack any sedimentary structures. In contrast the coarser grained beds are carbonate rich, the invertebrates are preserved as calcite (not aragonite) and the beds may show sedimentary structures, including symmetrical ripple lamination, crude fining upward and mud drapes. This author considers this cyclicity to be due to storm activity on a shallow marine shelf, in exceptional circumstances storm waves impinge on the sea floor beneath fair weather wave base. This causes the fine detrital material to go into suspension, stirring up, disassociating and abbrading the bioclastic debris and forming symmetrical ripples and mud drapes (Reif, 1982). Storms may be of variable intensity and bring in different proportions of reworked lithoclastic debris, possibly from different sources. As a result any individual storm may cause a variety of facies to develop depending on its provenance, including a shell bed (as in STA 5), or, where a coarse detrital fraction has been reworked, a sandy horizon (as in STA 3). Faunal variation has been recognised (Ivimey-Cook, 1974; Mayall, 1979; and this author) and related to salinity variations, although in the light of the preceding comments where the fauna is not in-situ, variation may reflect different source areas, rather than assuming, variable palaeoecological parameters.

Some diagenetic effects within the Westbury Formation are pronounced, notably the cone-in-cone formation (as seen in STA 3) and the decalcification of the original aragonitic pelecypods (as in STA 4). The reasons for these almost certainly relate to the long term anoxic condition that prevailed during deposition of the Westbury Formation.

A model for bacterial diagenesis in anoxic muds has been proposed by Curtis (1980). In the absence of oxygen, sulphate reducing bacteria thrive on the organic rich, detrital mud and produce bicarbonate ions, hydrogen and hydrosulphide ions. The sulphate reducing bacteria utilise dissolved sulphate diffusing through the sediment from the sea water, as an energy source.

The results of the chemical reactions that follow, depend on the amount of iron in the sediment. Where it is in abundance, iron pyrites will form, (as in STA 3). If it is

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relatively scarce, the dissolved sulphide levels build up, causing the pH to drop and the dissolution of any aragonitic pelecypod debris, (as in STA 4). Locally where pH conditions are more favourable, the bicarbonate ions and dissolved aragonite, will contribute to carbonate cement formation.

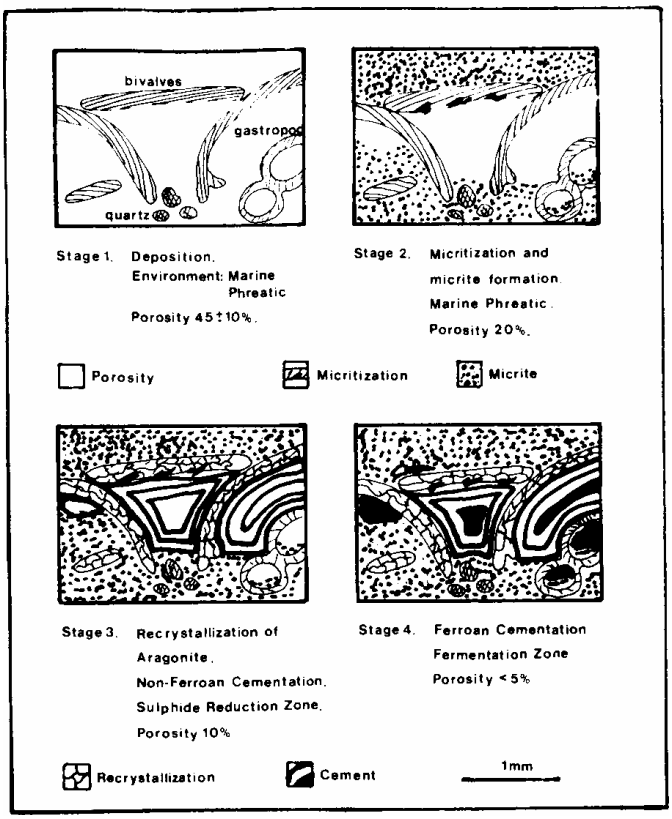

Figure 8. Idealized Westbury Formation diagenetic sequence (modified after Longman 1981).

In such areas a characteristic non-ferroan cement will form, because iron is preferentially incorporated into the pyrite (as in STA 3 and STA 5) rather than into an iron rich cement. A hypothetical cementation sequence can be erected (Fig. 8), which is controlled by the early sulphate reduction reaction and the presence of iron and results in early non-ferroan cements.

In localized areas of reduced pH phosphate will be concentrated in preference to calcite, thereby increasing the relative concentration of the vertebrate debris. This is then available for reworking by storm activity and has the potential of forming a "bone-bed". Thus the model for forming a "bone-bed" (as in STA 3) as a strand line lag (Hamilton, 1977; and Whittaker and Green, 1983) is rejected, in favour of a model involving an early diagenetically controlled submarine accumulation, which has subsequently been reworked and then cemented and therefore shows both diagenetic features and transportational features.

With the eventual depletion of dissolved sulphate in pore water, as a consequence of increasing burial depth, fermentation micro-organisms take over the processes of break down of organic rich muds. The products of this bacterially-catalysed reaction are methane, bicarbonate ions and hydrogen (Curtis, 1980). They then react with the detrital component, in particular any undissolved iron compounds to produce ferroan cements (as in STA. 2 and 3), (For typical Westbury Formation diagenetic stages; in particular late stage ferroan cements see Fig. 8).

The possibility of normal freshwater phreatic diagenesis, occurring (Longman, 1981), cannot be entirely ruled out, especially in the light of STA 5, where no micrite exists and only one non-ferroan cement zone is present.

Marshall (1982) interprets the final diagenetic alteration to be cone-in-cone calcite formation, associated with overpressuring and dewatering of the shales, as in STA 3.

Conclusions

i) The Westbury Formation is interpreted as having formed in predominantly shallow shelf marine conditions.
ii) Episodic interruption of low energy conditions by storms has caused the variable input of lithoclastic debris (depending on provenance), the formation of symmetrical ripple laminations and mud drapes and the reworking of the bioclastic material into discrete horizons. iii) Bacterial controlled diagenesis (Curtis, 1980) has mobilized carbonate forming decalcified horizons, forming characteristic non-ferroan and ferroan cements and with further reworking has the potential to form "bone-beds,.
iv) Westbury Formation sedimentary profiles are therefore a product of primary sedimentary features and early diagenetic modification.
v) Faunal variation is an artefact of transportation from different palaeoecological zones. vi) Formation of, micrites (STA 2) is a result of bacterially controlled diagenetic modification of organic rich black shales and is not a primary depositional feature.
vii) Cementation has occurred as a void-filling process in the coarser lithoclastic arenaceous horizons. viii) The cathodo-luminoscope is a very useful tool for distinguishing primary depositional fabrics from secondary diagenetic effects.

Acknowledgements This work was done at the University of · Bristol under the supervision of Prof. R.J.G. Savage, whilst in receipt of a Bristol University Scholarship. I am grateful to the Open University for allowing me to use their cathodo-luminoscope and to the following for useful discussion: A. Dickson, M. Coleman, D. Hamilton, W-E. Reif, H. Jenkyns and C. Jeans.

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Read at the Annual Conference of the Ussher Society, January 1984

The Blue Anchor Formation (late Triassic) in Somerset

G.WARRINGTON
A.WHITTAKER

Warrington, G. and Whittaker, A. 1984. The Blue Anchor Formation (late Triassic) in Somerset. *Proceedings of the Ussher Society, 6,* 100-107.

The Blue Anchor Formation comprises the highest beds of the Mercia Mudstone Group. The thickest, most fossiliferous, and best-exposed development is on the Somerset coast at and near the type locality, here designated, at Blue Anchor (ST03854368). There the formation comprises a succession of predominantly grey and green mudstones, silty mudstones and siltstones, with sulphate evaporites. These deposits accumulated largely in supra- and intertidal sabkha environments transitional between those represented by the underlying dominantly red-brown supratidal sabkha or playa deposits of the Mercia Mudstone Group, and the succeeding marine sediments of the Westbury Formation of the Penarth Group. The transitional nature of the formation is emphasised by the character of macro- and microfossil associations known from the type area. In the lower part of the sequence these comprise only miospores and trace fossils but in the higher beds precursors of the profuse bivalve and vertebrate faunas and organic-walled microplankton associations known from the Penarth Group are present and reflect the onset of marine conditions. Miospore assemblages of latest Triassic, late Norian(?) to Rhaetian, age are present throughout the formation in the type area.

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Introduction

The Blue Anchor Formation (Warrington et al., 1980, p. 55) is the highest formation in the Mercia Mudstone Group and comprises beds lying between the dominantly red-brown mudstones that constitute the greater part of that group, and the Westbury Formation of the succeeding Penarth Group. It encompasses the lithologically heterogeneous beds formerly termed 'Grey Marl', 'Grey and Tea Green Marl', 'Tea Green and Grey Marl', etc., and the lower part of the 'Sully Beds', of southern England and South Wales, and, elsewhere in Britain, the lithologically more uniform 'Tea Green Marl'. The Somerset coast around Watchet has been designated the type area of the formation, with representative sections at St. Audrie's Bay (ST105431) and Blue Anchor (ST040436) (Warrington et al., loc. cit.). The detailed lithostratigraphy of the formation in coast sections at and to the east of Watchet has been documented by Whittaker and Green (1983) who used the terms 'Tea Green Marl' and 'Grey Marl' for lower and upper divisions of the formation respectively; this terminology is adopted in the present account. The detailed stratigraphy of the formation in more westerly outcrops in the type area has not been documented since the time-of Richardson (1911), though sedimentological features of the sequence in the area were described by Mayall (1981) who divided the formation into a Rydon Member and a younger, impersistent, Williton Member. The detailed lithostratigraphy of the Blue Anchor Formation at the type locality, here designated, at Blue Anchor (Fig. 2) was recorded by one of us (A.W.) during the resurvey of the Weston-super-Mare district (1:50 000 geological sheet 279; Whittaker and Green 1983) and parts of contiguous districts. It is presented here in a style compatible with that used by Whittaker and Green (op. cit., pp. 49-53) for sections at Watchet, St. Audries's Bay and Lilstock, and with supplementary notes on outcrops between Blue Anchor and Watchet, to extend the lithostratigraphic documentation of the formation to the western end of the type area in a standard format.

Other occurrences of the formation in Somerset (Fig. 1) and the macropalaeontology of the unit in that area are reviewed. The micropalaeontology of the section at St. Audrie's Bay is documented, and the age and mode of origin of the formation in Somerset are considered.

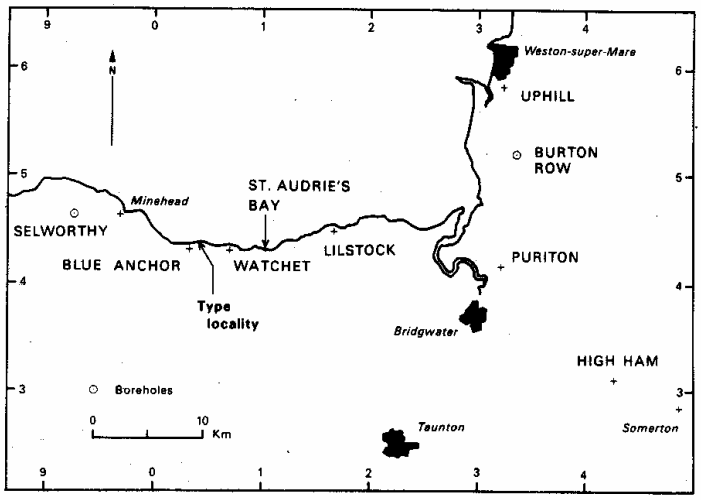

Figure 1 Localities in west Somerset.

The type section and other occurrences of the Blue Anchor Formation on the Somerset coast west of Watchet

The Blue Anchor Formation is well exposed in cliffs and on the foreshore between Watchet and Blue Anchor; the most westerly exposures in this area comprise poorly visible south-south-westerly dipping reefs on the foreshore (ST02454352).

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The formation is exposed in the core of an anticline visible in the cliff and foreshore at Blue Anchor Point (ST03654390; Bradshaw and Hamilton 1967; Hamilton and Whittaker 1977) where a fauna including bivalves and vertebrate remains was recovered from an horizon about 3.05 m below the Westbury Formation bone bed on the foreshore (Boyd-Dawkins 1864a, b). The section of the formation from this site (Fig. 2) is here designated the type section (Table 1); it is compiled from measurements made in the cliff face at ST03854368.

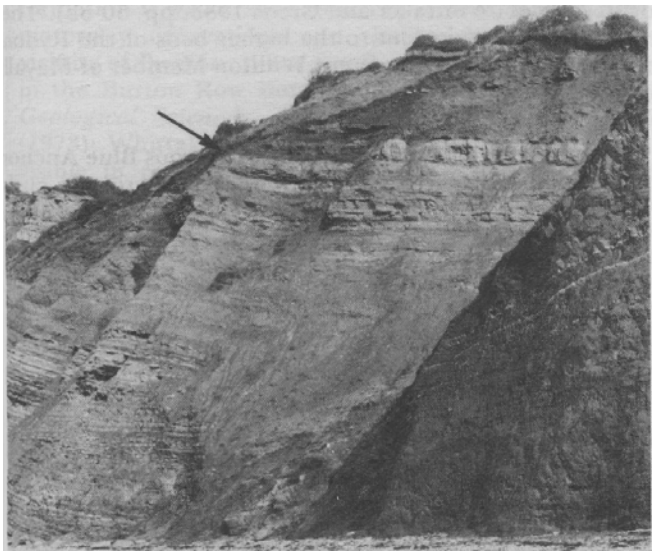

Figure 2 The type locality of the Blue Anchor Formation at Blue Anchor Cliff, Somerset, in 1972 (British Geological Survey photograph A.11715). The cliff exposure, between the arrow and beach level, comprises Grey Marl of the type section; lower beds, and the base of the formation, occur off the photograph to the left.

Table 1. Type section of the Blue Anchor Formation at Blue Anchor Cliff (ST03854368); thicknesses in metres.

Penarth Group

WESTBURY FORMATION

Mercia Mudstone Group

BLUE ANCHOR FORMATION

Grey marl (c.32.2 m):

Mudstone, dark grey, shaly and laminated:	0.13
Siltstone. bullish grey, argillaceous and blocky:	0.10
Siltstone yellowish buff. argillaceous and blocky:	0.28
Siltstone, greenish buff. blocky:	0.33
Siltstone, greyish buff, with beds of dark grey mudstone:	0.36
Mudstone, dark grey:	0.23
Mudstone, greenish grey, fine-grained, blocky:	0.23
Mudstone. greenish grey, coarse-grained, blocky:	0.43
Mudstone. dark grey, shaly:	0.69
Mudstone. greenish grey, coarse-grained, blocky:	0.18
Mudstone, dark grey, shales:	0.13
Mudstone, greenish grey, coarse-grained and blocky:	0.23 - 0.30
Mudstone, dark grey and shaly:	0.23
Mudstone, greenish grey, blocky:	0.10
Mudstone, dark grey, shaly:	0.15
Mudstone, mainly buffish grey, hard, blocky and silty, with some shaly greenish grey beds up to 0.08 m thick:	0.81
Mudstone, olive green and buff, fine-grained and blocky:	0.46
Mudstone, mainly dark grey and shaly but with some alternations of buff and olive green marls:	1.52
Mudstone, olive green and buff, silty and with abundant angular gypsum crystals:	0.08
Mudstone, dark grey, shaly and with some thin, lighter grey mudstone beds:	1.09
Mudstone, greenish grey, coarse-grained and blocky:	0.05
Mudstone, dark greenish grey:	0.61
Mudstone, greenish grey and buff, laminated and silty:	0.61
Mudstone, grey, blocky, medium to coarse-grained:	0.25
Mudstone, grey, blocky, coarse-grained and silty:	0.30
Mudstone, dark grey:	0.30
Mudstone, pale greyish green, blocky and coarse-grained:	0.15
Mudstone, greyish green and blocky:	0.41
Mudstones, alternating pale and dark green, with gypsum veins:	0.66
C - Mudstones, alternating pale and dark grey and greyish green with abundant bedded nodular gypsum and veins. Individual mudstone beds are between 0.08 m and 0.13 m thick:	2.74
Mudstone, greyish green with abundant gypsum:	0.25
Mudstone, grey, with gypsum:	0.43
Mudstone, greyish green, blocky and coarse-grained:	0.23
Mudstone, alternating light grey and dark grey:	0.33
Gypsum, two veins, very persistent laterally:	0.05 - 0.08
Mudstone, greyish green, blocky and with gypsum veins:	0.91
Mudstones, alternating light and dark green, in 0.02 m beds:	0.56
Gypsum, nodular and bedded:	0.08 - 0.15 m
Mudstones, alternating greyish green and dark grey in 0.15 m beds. with gypsum veins:	0.51
Gypsum:	0.05 - 0.15
Mudstones. alternating greyish green and dark grey beds:	0.23
Mudstone. grey and green, with abundant gypsum in the lowest 0.23 m:	0.66
Mudstones. alternating greyish green and dark grey beds:	0.61
Mudstone, green, blocky, with much gypsum:	0.38
Mudstone dark grey and greenish grey:	0.34
B- Mudstone, greenish grey, With gypsum nodules up to 0.10m:	0.30
Mudstones, green and grey:	0.38
Mudstone. greenish grey, blocky and with pink gypsum nodules and white veins:	0.36
Siltstone, greenish grey, wispy bedded:	0.23
Mudstone. dark grey:	0.13
Siltstone. greenish grey, wispy bedded with	0.05
0.08 m gypsum bed 0.08 m below the top:	0.53
Mudstones dark grey and green alternating, with ramifying gypsum veins and occasional bedded. nodular gypsum:	2.24
Mudstone, greyish green, silty and blocky, with some thin beds of green laminated mudstone:	0.53
Mudstone. greenish grey and dark grey, blocky:	0.56
Mudstone, greyish green, silty, hard, with gypsum nodules:	0.86
Mudstones, greyish green and dark grey in alternating	

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beds up to 0.18 m thick. Occasional beds of nodular gypsum: 2.29
Gypsum: 0.08
Mudstones, greyish green and dark grey in alternating beds up to 0.08 m thick: 0.71
A-Gypsum: 0.05
Mudstone green, coarse-grained and blocky, with a 0.08 m wispy-bedded unit 0.23 m below the top: 0.69
Mudstones, alternating greyish green and dark grey beds up to 0.28 m thick, with abundant white and pink gypsum veins: 2.67

Tea Green Marl (4.34 m):

Siltstone. greyish green, hard. wispy-bedded and poorly laminated: 0.15
Mudstone. green, with white gypsum: 0.30
Siltstone. greenish grey, hard, wispy-bedded: 0.30
Mudstone, green, with gypsum nodules: 0.46
Mudstone, grey and green, with gypsum in the topmost 0.30 m: 0.41
Mudstone, green, with some laminated beds: 1.14
Gypsum: 0.02
Mudstone, green, with two red speckled beds in the middle: 0.25
Siltstone, greenish grey, hard and wispy-bedded: 0.10
Mudstone. dark greenish grey: 0.03
Mudstone, green: 0.15
Siltstone greenish grey, hard. laminated and wispy-bedded in the upper part, pebbly in the lower: 0.18
Mudstone. green, with gypsum nodules: 0.25
Siltstone. greyish green hard and with laminae, Gypsum nodules: 0.30
Mudstone, green, blocky and with purplish red bands: 0.30
(base of Blue Anchor Formation)
undifferentiated beds in the Mercia Mudstone Group Succession.

The lower division of the Blue Anchor Formation, the Tea Green Marl, comprises green and dark greenish grey silty mudstones or siltstones; its base is placed at the top of the highest prominent bed of red mudstone in the Mercia Mudstone Group and its top is marked by the base of the lowest dark grey mudstone bed within the formation (Whittaker and Green 1983, p. 39). It is 4.34 m thick at the type locality and thickens eastwards to 5.18 m at St. Audrie's bay. The Tea Green Marl is equivalent to the basal part of the Rydon Member of Mayall (1981).

The base of the Grey Marl, the upper division of the Blue Anchor Formation. is placed at the base of the lowest dark grey mudstone bed within the formation' the top of the unit. and the top of the formation, is marked by the base of the succeeding Westbury Formation (Whittaker and Green 1983, p. 39). The Grey Marl is some 32.2 m thick in the type section but thins eastwards to about 26.0 m at St. Audrie's Bay; it comprises alternating dark grey mudstones, some of which are shaly, and greenish grey or buffish grey silty mudstones which are dolomitic in part. Finely laminated and, in places, burrowed mudstones occur and small-scale collapse structures, possibly caused by solution of evaporite minerals, are present. Veins and nodular masses of colourless, white and red-tinged gypsum occur throughout the Grey Marl at and near the type locality, but are abundant only in beds below 10.67 m from its top. Units which are particularly rich in nodular gypsum at Blue Anchor are designated B and C on the section (Table 1); these correspond approximately with Richardson's (1911) beds 12 and 13, and 6 to 8 respectively and are traceable in other sections of the Grey Marl farther east on the Somerset coast. Another laterally extensive and correlatable unit, designated A in the type section (Table 1), occurs in the lowest 5 m of the Grey Marl at a level approximately equivalent to bed 18 of Richardson (1911). Elsewhere, a thin bed of red marl associated with gypsum is present at about the level of unit A and overlies a hard grey siltstone bed, commonly with cavities (Whittaker and Green 1983 pp. 50-52). The Grey Marl is equivalent to the higher beds of the Rydon Member and the impersistent Williton Member of Mayall (1981).

Eastwards from the type locality, gypsiferous Blue Anchor Formation strata are exposed in faulted outcrops in cliffs and on the foreshore (Whittaker 1972, P1.12; Hamilton and Whittaker 1977, figs 1 and 3). A faulted outlier (ST05274365) of Blue Anchor Formation beds occurs in a syncline offshore from Warren Farm, and steeply-dipping gypsiferous beds of this formation are seen in the nearby cliff (ST05364343). An excellent, though largely inaccessible, section through the formation is present in a cliff (ST0600433t) about 850 m east of Warren Farm. The stratigraphic levels and thicknesses of the prominent gypsum-bearing units at this locality were estimated by suspending a scale from the cliff top and photographing the section from beach level. The bases of units B and C are about 22.25 m and 10.68 m below the top of the formation respectively; unit B is about 1.83 m thick and unit C is about 2.74 m thick. Unit A occurs about 28.35 m below the top of the formation.

About one kilometre west of the northern breakwater of Watchet harbour, the Blue Anchor Formation is exposed in the core of an east-west-trending pericline (c. ST063437; Whittaker 1972; Hamilton and Whittaker loc. cit.). The formation is faulted on the foreshore to seaward from Watchet Harbour (ST073437) but is traceable, in a slightly flexured but continuous outcrop, eastwards to Helwell Bay (ST080434) where it is faulted against Lower Lias strata (Hamilton and Whittaker 1977, fig. 3).

The Blue Anchor Formation elsewhere in Somerset

The most westerly occurrences of the Blue Anchor Formation preserved in Somerset are near Selworthy (Fig. 1), on the southern slopes of North Hill, west of Minehead. There the formation is poorly exposed, but the beds strike east-south-eastwards from Holnicote (SS914463) to Venniford Cross (SS932458) and dip at up to 23° to the north-north-east. The formation was proved between 35.61 and 51.44 m in the Selworthy No. 2 Borehole (SS92444618; *Report Series, Institute of Geological Sciences*, No. 74/7, p. 2 (1974); Whittaker 1976). The 15.83 m of the formation proved in this borehole is in faulted contact with disturbed red marls and the full thickness in the Selworthy area is, therefore, unknown. The lithological character of the formation in the borehole is similar to that seen elsewhere in Somerset; predominantly grey and greenish

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grey siltstones alternate with dark grey or black mudstones which are slightly disturbed in places and have listric surfaces. Veinlets and thin (c.0.01 m thick) layers of anhydrite occur in the lowest 3 m of the Grey marl which rests upon a possible representative of the Tea Green Marl, seen for 0.08 m only.

Coastal sections and other exposures and occurrences of the Blue Anchor Formation to the east of Watchet have been documented by Whittaker and Green (1983); the formation is well-known from the records of railway and motorway excavations and shallow boreholes in the vicinity of Puriton, Brent Knoll and Uphill. It was proved in the Burton Row 13orehole, Brent Knoll (*Institute of Geological Sciences, Annual Report for 1972*, p. 122 (1973); Whittaker and Green 1983, pp. 121-123), and is visible in road sections at Uphill (ST325582). To the south-east of Puriton the formation has been recorded in site investigation boreholes in the High Ham area near Somerton; there it is c. 19 m thick and similar in lithology to the occurrences in the type area father west. However, in addition to being thinner and containing only minor occurrences of gypsum in thin sub-horizontal veins, the High Ham sequence is notable for the presence of a pebbly bed about 5.5 m below the top of the formation. The Tea Green Marl constitutes about 3.0 to 3.5 m of the Blue Anchor Formation succession in this area. Farther east, in the Bruton No. 1 Borehole (ST68963284), some 16 km south-west of Frome, the formation comprises 6.6 m of green to grey mudstones with traces of gypsum and with interbedded grey to green calcareous siltst0nes and limestones (Holloway and Chadwick 1984).

Palaeontology of the Blue Anchor Formation in Somerset

The Blue Anchor Formation in Somerset has yielded trace fossils and small but varied assemblages of macrofossils, including indeterminate plant debris, molluscan and vertebrate remains, and microfossil associations comprising miospores and organic-walled microplankton.

Trace fossils

Boyd-Dawkins (1964a, pp. 39%398) reported "very numerous holes and tracks" between 2.44 and 4.27 m below the top of the formation and observed "casts of Fucoids and trails of Annelids" in the top 1.83 m of the unit near the type locality. Burrows occur in the top 1.52 m of the formation at St. Audrie's Bay (Bristow and Etheridge 1873; Richardson 1911, p. 24; Whittaker and Green 1983, p. 50). The presence of *Arenicolites* and other trace fossils in the topmost beds of the formation on the west Somerset coast was noted by Stevenson and Warrington (1971); Mayall (1981) observed that only *Arenicolites* occurs in the lower part of the formation (Rydon Member) but that *Diplocraterion* burrows penetrate the upper surface of that unit and occur, with Arenicolites, *Muensteria*, *Planolites*, *Rhizocorallium* and *Siphonites*, in the overlying beds (Williton Member) near Blue Anchor.

Macrofossils

Bivalves from about 3.05 m below the top of the formation near the type locality (Boyd-Dawkins 1864a, pp. 397-398) were identified by Etheridge as "*Modiola minima*, *Pecten valoniensis*, *Myacites striatogranulata*, *Gervillia praecursor*, *Pullastra arenicola* and *Cardium rhaeticum"; P. arenicola* was also recorded from the top 1.83 m of the succession (Boyd-Dawkins, *loc. cit.*). At St. Audrie's Bay, *Gervillia praecursor* occurs within the top 1.52 m of the succession (Bristow and Etheridge 1873; Richardson 1911, p. 24; Whittaker and Green 1983, p. 51) and *Mytilus*? was recorded from 1.74 to 2.20 m lower in the formation (Whittaker and Green 1983, p. 50). Farther east, at Lilstock (c. ST176453), Richardson (1911, p. 30) recorded "*Pteria contorta* very common" associated with "*Volacilla sp.*" and much shell debris from "bed 1", at the top of his 'Sully Beds' unit; this bed is now considered part of the Westbury Formation (Whittaker and Green 1983, p. 52) and its fossils, are, therefore, excluded from the fauna of the Blue Anchor Formation. A similar consideration may apply to the record (Richardson 1911, p. 35) of "*Pteria contorta*" from a comparable horizon in the succession at Dunball (c.ST313412). Beds 0.70 to 1.58 m below "bed 1" at Lilstock (Richardson 1911, p. 30) yielded "*Pteria*" and *Protocardia*?; "*Pteria contorta*" was recorded immediately below "bed 1" and shell debris was noted between 1.98 and 2.69 m below the top of the formation at this locality (Richardson, *loc. cit.*).

Mayall (1981) noted that marine bivalves do not occur in the lower beds (Rydon Member) of the formation but that generally poorly preserved specimens of *Chlamys sp.*, *"Eotrapezium sp."*, *Gervillia praecursor*, *Modiolus sp.* and *Protocardia sp.* are commonly seen on bedding surfaces in the overlying beds (Williton Member) at Blue Anchor.

A gastropod (*Chemnitzia* or *Turritella*) and a cephalopod ("allied to *Beloteuthis* or *Geoteuthis*") were reported by Boyd-Dawkins (1864a, p. 398) from 2.44 to 4.27 m below the top of the formation near Blue Anchor but were omitted from a subsequent account (Boyd-Dawkins 1864b, p. 260). Fragments of gastropods occur at a similar horizon at St. Audrie's Bay (Whittaker and Green 1983, p. 50).

Remains of the fish *'Sargodon tomicus'* and *Saurichthys apicalis* occur about 3.05 m below the top of the formation near Blue Anchor; *Acrodus minimus*, *Gyrolepis alberti* and *G. tenuistriahtus* are recorded from the same level and also within the top 1.83 m of the succession (Boyd-Dawkins 1864a, b). Unidentified fish remains were reported from c.4.27 m below the top of the formation in the same area (Richardson 1911, p. 18). At St. Audrie's Bay, *Gyrolepis* and *Hybodus* cf. *cloacinus* occur associated with teeth of "*Sargodon*" and "*Sphaerodus"* type 3.26 to 3.72 m below the top of the formation (Whittaker and Green 1983, p. 50) and fish scales are present within the top 1.52 m of the succession (Bristow and Etheridge 1873; Whittaker and Green 1983, p. 51). At Lilstock, fish scales were noted to 2.41 m below the top of the formation (Richardson 1911, p. 30). The topmost bed ("bed 1") of the "Sully Beds" at this locality (Richardson 1911, p. 30), which yielded *Gyrolepis alberti*, *Lepidotes*? and *'Sargodon tomicus'*, is now regarded as part of the Westbury

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Formation (Whittaker and Green 1983, p. 52).

Rolled fragments of bone, regarded as reptilian in Origin, were recorded from 2.44 to 4.27 m below the top of the formation near Blue Anchor (Boyd-Dawkins 1864a, p. 398). A small plesiosaurian tooth from the top of the "Sully Beds" at Lilstock (Richardson 1911, p. 30) originated from a bed now regarded as part of the Westbury Formation (Whittaker and Green 1983, p. 52). A tooth, regarded as mammalian in origin, was described and named *Hypsiprymnopsis rhaeticus* by Boyd-Dawkins (1864a, pp. 409-12); it was redescribed and transferred to the genus *Microlestes* by Owen (1871). More recent workers (Clemens *et al.*, 1979, p. 13) consider that the specimen, from an horizon 3.20 m below the top of the formation near Blue Anchor (c. ST045438) and now apparently lost, warranted no more refined a determination than "?tritylodontid, *incertae sedis*".

Other remains recorded from the formation at Blue Anchor include fragments of wood from 2.43 to 4.27 m below the top of the unit (Boyd-Dawkins 1864a, p. 398) and a small coprolite from about 4.27 m below the top (Richardson 1911, p. 18).

Microfossils

Palynomorph assemblages recovered from the highest beds of the Mercia Mudstone Group, including the Blue Anchor Formation, at St. Audrie's Bay and in the Burton Row Borehole, Somerset (Warrington 1974, 1978a, 1980, 1981; pp. 131-2 in Whittaker and Green 1983) currently constitute the most extensive palynological documentation of those beds in Britain. The palynomorph succession from the Blue Anchor Formation at outcrop at St. Audrie's Bay is illustrated in Figure 3 with, for comparison, assemblages recovered from the underlying beds in the Mercia Mudstone Group and from the lowest beds of the succeeding Westbury Formation (Penarth Group) exposed at that locality.

Organic residues from 29 levels in the c. 31.18 m thick Blue Anchor Formation succession at St. Audrie's Bay have been examined for palynomorphs. The Tea Green Marl unit, sampled at six levels, proved unproductive but 12 of the remaining 23 samples, from the overlying Grey Marl unit, yielded palynomorphs; the majority of the productive samples were from the upper half of the succession (Fig. 3) and the two highest are from beds equivalent to the Williton Member of Mayall (1981). The majority of the assemblages comprise only miospores but organic-walled microplankton occur in the highest sample.

The miospore associations are of limited diversity and relatively uniform composition (Fig. 3); they are dominated by circumpolles, principally *Classopollis torosus* (Reissinger) Balme 1957 but including *Granuloperculatipollis rudis* Venkatachala and Góczán *emend*. Morbey 1975 and small numbers of *Corollina zwolinskai* Lund 1977 and *Gliscopollis meyeriana* (Klaus) Venkatachala 1966. Other typical components are *Leptolepidites argenteaeformis* (Bolchovitina) Morbey 1975, *Ovalipollis pseudoalatus* (Thiergart) Schuurman 1976, *Rhaetipollis germanicus* Schulz 1967, and bisaccate taxa including *Alisporites spp.* and *Vesicaspora fuscus* (Pautsch) Morbey 1975. Associations from the lower 11-5 m of the formation are comparable in composition with the few recovered from beds in the Mercia Mudstone Group exposed beneath the Blue Anchor Formation at St. Audrie's Bay. A slight increase in diversity is apparent in associations from the top 10 m of the formation where *Acanthotriletes ovalis* Nilsson 1958, *Carnisporites spp.*, *Kraeuselisporties reissingeri* (Harris) Morbey 1975, *Ricciisporites tuberculatus* Lundblad 1954 and *Quadraeculina anellaeformis* Maljavkina 1949 are present in small numbers. These taxa are more typical of assemblages from the Penarth Group, or younger beds, and the miospore associations from the higher part of the Blue Anchor Formation thus have affinities with and are transitional to those of the overlying Westbury Formation (Fig. 3). A notable feature of palynomorph assemblages from that formation is the commonly abundant presence of organic-walled microplankton associations which are typically dominated by the dinoflagellate cyst *Rhaetogonyaulax rhaetica* (Sarjeant) Loeblich and Loeblich *emend*. Harland et al., 1975. The sporadic occurrence of specimens of *R. rhaetica* in the highest beds of the Blue Anchor Formation, equivalent to the Williton Member, at St. Audrie's Bay, and of that taxon and other organic-walled microplankton in beds at a similar level in the sequence proved at Brent Knoll (Warrington 1980, 1981) and elsewhere, is a precursor of their widespread importance in assemblages from the succeeding Penarth Group (vide: Orbell 1973; Warrington 1977a, b, 1978b, 1982).

Conditions of deposition

The conditions of deposition of the Blue Anchor Formation in west Somerset have been discussed by Stevenson (1970), Mayall (1981) and Whittaker and Green (1983).

Lithostratigraphic evidence

The lower beds in the formation (Tea Green Marl) contain burrows, some laminated units, and local intraformational unconformities which herald the more general development of those features in the higher beds (Grey Marl). Of particular interest in the Grey Marl is the association of porphyroblastic sulphate nodules with carbonaceous mudstones of possible algal mat origin and suggestive of a supratidal origin in a sabkha environment (*vide*: Stevenson 1970; Sellwood et al., 1970; Stevenson and Warrington 1971). Alternating with these sediments are various lithologies, including laminated dolomitised

Figure 3 Palynomorph assemblages from the Blue Anchor Formation and contiguous beds exposed at St. Audrie's Bay, Somerset. Relative abundances of miospore and organic-walled microplankton taxa and other remains (A), of organic-walled microplankton (black) to other palynomorphs (B), and of major groups of palynomorphs (C), are expressed as percentages based upon counts of 200 specimens. Preparations and slides are held in the micropalaeontological collections at the British Geological Survey, Keyworth. Nottingham and are registered in the SAL series

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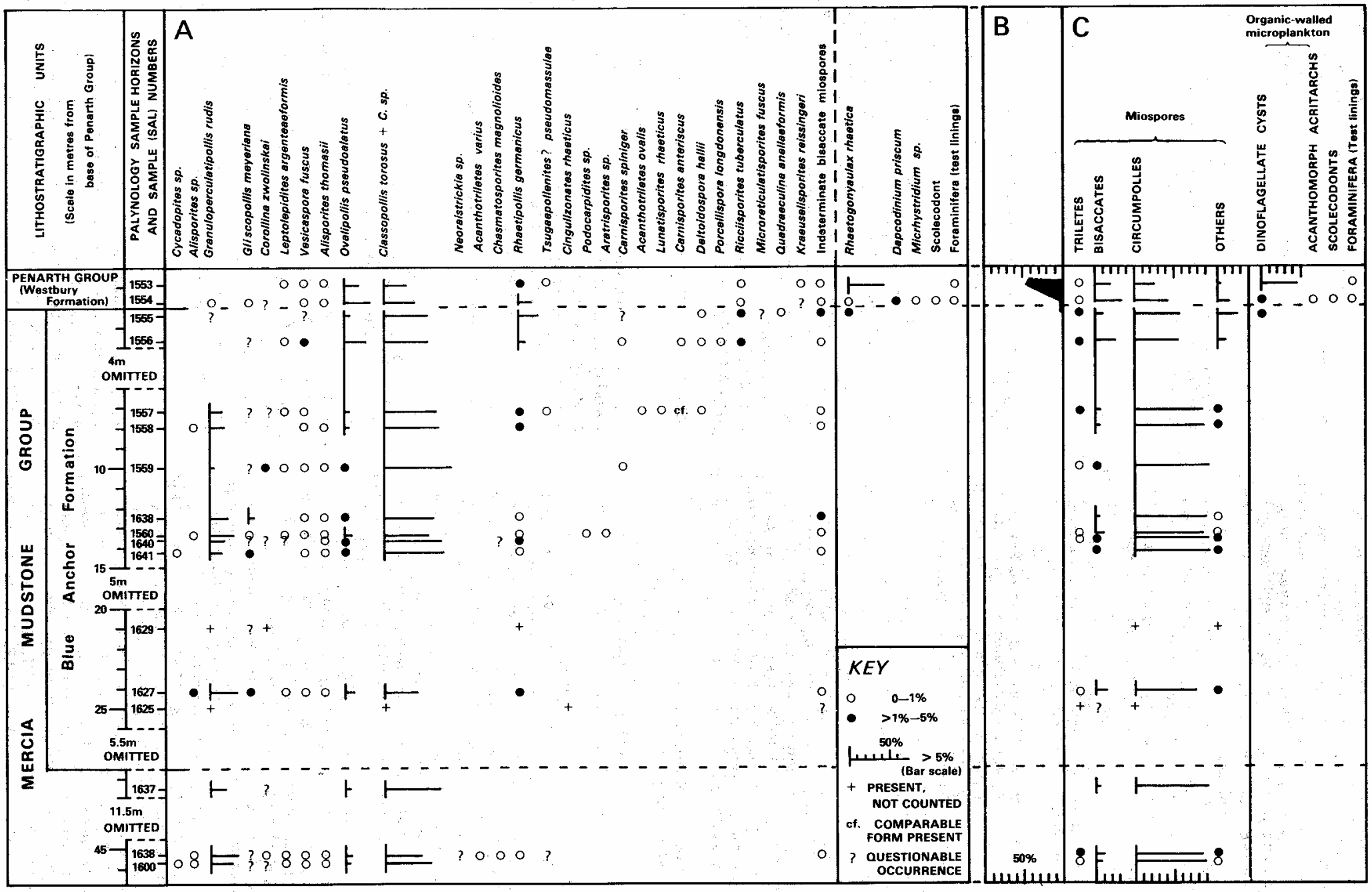

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siltstones, which are indicative of a low-energy intertidal depositional environment and which, in some instances, contain marine fossils. The lithostratigraphic succession in the Blue Anchor Formation thus reflects numerous advances and retreats of an aqueous environment and the transition to one of marine character near the top of the sequence.

Palaeontological evidence

The majority of the palynomorph assemblages from the Blue Anchor Formation comprise only remains of land plant origin and are dominated by miospores (e.g. *Classopollis*; Fig. 3) of coniferalean origin. The appearance of small numbers of organic-walled microplankton in assemblages from the highest beds in the succession indicates ,that those beds formed in an aqueous environment of marine origin. A comparable indication is afforded by the stratigraphic distribution of the bivalves and fish remains which are the principal constituents of the sparse macrofaunas known from the formation.

The character of these macro and microfossil associations emphasises the transitional nature of the unit which comprises deposits reflecting the change from an essentially terrestrial environment, represented by the lower part of the formation and underlying beds in the Mercia Mudstone Group, to one increasingly influenced by a transgressive marine environment in which deposits forming the higher part of the formation and the succeeding Penarth Group accumulated.

Age of the Blue Anchor Formation in Somerset

A latest Triassic age is assigned to the formation on the basis of the bivalve taxa present and also from the stratigraphically more extensive palynomorph assemblages. The presence of *Rhaetipollis germanicus* throughout the formation, and also in underlying beds in the Mercia Mudstone Group succession (Fig. 3), is, by comparison with latest Triassic sequences elsewhere, indicative of a late Norian (?) to Rhaetian age (Morbey 1975, 1978; Schuurman 1977, 1979; Fisher and Dunay 1981; Visscher and Brugman 1981; Smith 1982; Orlowska-Zwolinska 1983).

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Read at the Annual Conference of the Ussher Society, January 1984

Uranium in the New Red Sandstone of south-east Devon

E.M. DURRANCE

Durrance, E. M. 1984. Uranium in the New Red Sandstone of south-east Devon. *Proceedings of the Ussher Society, 6,* 108-115.

The results of a stream water survey in south-east Devon of uranium in solution, show three north-south trending belts of high values. These belts follow the north-south trending outcrop of the New Red Sandstone Series (Permian-Triassic). The westerly belt follows the outcrop of the Permian breccias north of Torbay, the central belt extends inland from Littleham Cove on the Littleham Mudstone Formation and the easterly belt is coincident with the lower part of the Mercia Mudstone Group north of Sidmouth.

Large concretionary uranium-bearing nodules are found in outcrops of the Littleham Mudstone Formation and in the lower-part of the Mercia Mudstone Group. Sporadic occurrences of nodules also characterise the rest of the Mercia Mudstone Group. High stream water uranium values are thus interpreted as indicating the extent of uranium mineralisation within these formations and, additionally, similar mineralisation in the breccias.

The occurrence of uranium (and other rare element concentrations) in relatively impermeable lithologies, with the main mineralisation having taken place adjacent to the margins of the major aquifer of the New Red Sandstone (which lies between the Littleham Mudstone Formation and the Mercia Mudstone Group), is taken as indicative of the mineralisation processes involved. A mechanism of charging the aquifer by thermal groundwaters carrying some elements in solution, rising up major fractures which extend into basement Devonian and Carboniferous rocks, is proposed. Slow discharge from the aquifer (and minor aquifers elsewhere in the succession) then led to the formation of the deposits at suitable precipitation sites, such as organic debris. Mineralisation could be still continuing.

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Introduction

In south-west England uranium concentrations are found as vein mineralisation (Ball et al., 1979) and stratabound mineralisation (Harrison, 1975). Vein mineralisation of uranium usually occurs near the margins of the granite plutons, but secondary uranium minerals also occur deposited along fractures within the granites. In general the vein deposits are either part of the east-west trending main belt of mineralisation where the uranium is in the form of pitchblende, accompanying tin and copper; or occupying north-south cross-courses where again the uranium may be as pitchblende, but coffinite also occurs and the association is with nickel, cobalt, bismuth, iron, lead and zinc. Secondary uranium-bearing minerals are found associated with the zones of oxidised enrichment in both east-west veins and the cross-courses. These are typically, torbernite, autunite and zippeite.

In south-east Devon the main uranium mineralisation occurs in the form of dark concretionary nodules surrounded by pale green haloes in the otherwise red rocks of the Littleham Mudstone Formation of the New Red Sandstone near Exmouth. More dispersed uranium enrichment also occurs along the contact between green-coloured strata in the mudstones and the normal red mudstones. In this environment the uranium is associated with high levels of vanadium, copper, lead, zinc, nickel and cobalt.

Both the vein mineralisation and the concretionary nodules represent the migration of groundwater which carried the appropriate elements in solution through the rocks. That such migration occurred as more than one event is shown by radiometric dating of uranium minerals from the vein deposits of Devon and Cornwall and paragenetic relationships. Unfortunately radiometric age dating of the concretionary nodules of south-east Devon has been unsuccessful (R. K. Harrison, personal communication), probably because of disequilibrium in the uranium decay series (Durrance et al., 1980)

The origin of the uranium mineralisation associated with the granites has been investigated by Simpson et al., (1978). These authors concluded that deeply circulating meteoric waters could have leached uranium (and other elements) from the hot granite magma, or from hot granite (heat being needed to drive the convective circulation and allow uranium in resistate minerals to be leached), and given rise to precipitation in fractures on cooling. In contrast, Barbier (1974) thought that continental weathering (in the Permian) and mobilisation by low-temperature meteoric waters could account for the formation of the vein mineralisation. Durrance et al., (1982) considered that the circulation of groundwater, which was principally of meteoric origin, was driven by radiogenic heat from the granites. This heat source was supplemented periodically by enhanced mantle heat flow from hot-spots beneath south-west England. Fracture permeability was also periodically enhanced by tectonic activity. The operation of these factors could thus have caused the episodic reactivation of the system at different temperatures and under widely differing stress regimes from the late Carboniferous to the present day.

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Harrison (1975) also considered that introduction of the heavy metals in the concretionary nodules of south-east Devon had occurred from outside the New Red Sandstone sedimentary succession. This implies that the circulation systems were not confined to just the granites and their immediate vicinity.

Uranium Decay Series and Groundwater Movement

Oxygenated groundwater, which must be principally of meteoric origin, will readily leach and transport uranium, if it can transform U4+ to U6+ and if the Eh and pH conditions are such as to permit the formation of soluble complex ions. Contact .times required to produce a given concentration of uranium in the groundwater for any given groundwater chemistry will, however, also vary with the nature of the sites the uranium occupies in the host rock. In studies of the Dartmoor granite, Heath (1982) has even shown that uranium loss during weathering is more controlled by the site of the uranium in the rock than by the degree of weathering. Where only primary, detrital, uranium-bearing minerals are therefore present in sedimentary rock sequences, such as those in south-east Devon, long residence times are needed to increase the uranium content of young groundwater to appreciable levels. Thus, in south-east Devon, groundwaters that have been in contact with rocks in which the uranium (although at levels of about 4ppm) is not easily leachable, have uranium contents of less than lppb because of the very short residence time shown by their tritium values (Walton, 1982). Conversely, where oxygenated groundwaters pass through rocks where there is an occurrence of secondary uranium, the uranium levels in the groundwaters may rapidly rise to as much as 30ppb, depending upon the detailed geochemical environment that is present. Noting that uranium has a very long half-life, unless it is possible to date groundwaters by their tritium or helium content (Andrews, 1982), or to observe the presence of Secondary uranium minerals in the rock, it is unfortunately often impossible to resolve the interaction of time and availability. However, direct observations of secondary uranium minerals can often be made fairly easily, while high uranium levels in surface waters (even if these are flowing over rocks carrying secondary uranium)can indicate the discharge of uranium-rich groundwaters, especially if the surface waters maintain a high uranium concentration after periods of appreciable rainfall.

While residence times for surface waters in areas of secondary uranium deposits are large enough to produce significantly high levels of uranium in solution, that is not the case for ^{222}Rn. ^{222}Rn is a gaseous unstable daughter product of the decay of ^{238}U, and has a half-life of about 3.832 days. The contact time needed to allow the level of ^{222}Rn to increase to measurable levels in running surface waters is impractical; degassing and loss of radon is likely to take place as rapidly as. acceptance from any detrital radium in the stream sediment (Andrews and Wood, 1972). It is clear that the production of high radon levels in water also requires long residence times. Moreover, migration of radon is far more effectively controlled by movement of groundwater than by diffusion of the gas (Tanner, 1964). High radon levels in stream water therefore may be shown to be very largely controlled by the discharge of groundwaters passing through uranium-bearing rocks. The occurrence of the uranium in resistate minerals or as secondary minerals does not seem to be a sensitive control on radon levels, nor does the quantity of uranium that is present; the main control appears to be the residence time of the groundwater and the rate of rise · of the groundwater. The stream water radon anomaly of south-east Devon (Durrance, 1978) is therefore not related to uranium distribution (Durrance and Heath, 1985).

Uranium Concentrations in south-east Devon

The presence of uranium in reasonable quantities in parts of east Devon was first recognised by Carter (1931), who described large nodules from the red mudstones and sandstones of Littleham Cove, at the western end of Budleigh Salterton beach, which would produce contact autoradiographs. These nodules have since been subject to detailed study by Perutz (1939), Harrison (1962, 1975) and Durrance and George (1976). The occurrence of the nodules is principally in the coastal area of outcrop of the Littleham Mudstone Formation (Henson 1971), but some nodules have also been recorded from the brickworks quarry at Withycombe Raleigh. Small dark nodules ("fish eyes") also occur in the Mercia Mudstones.

Apart from the concentration of uranium in the nodules of the Littleham Mudstone Formation, it would appear that small amounts of uranium may be found throughout the area of the New Red Sandstone Series (Cosgrove, 1973). The geology of the area is shown in Figure 1.

Hydrogeochemistry

For uranium to be mobile in a particular pH range the appropriate complexing agent must be available.

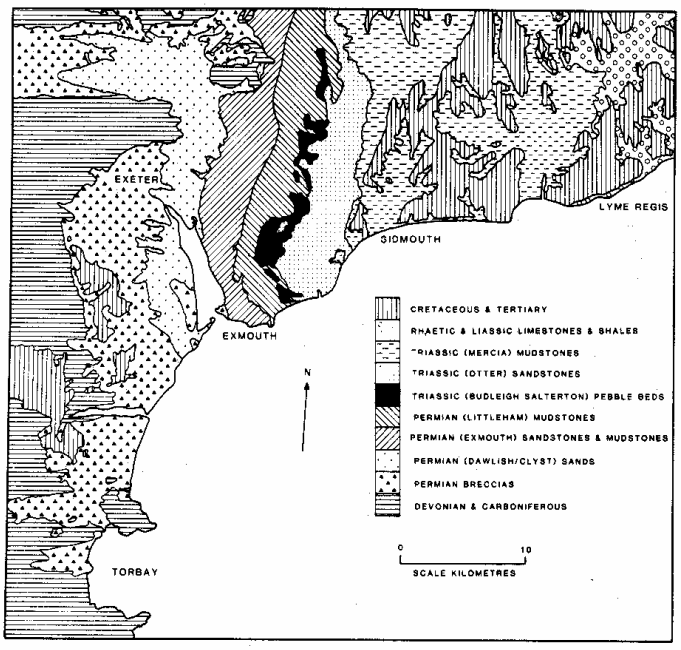

Figure 1. Geology of south-east Devon.

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Generally for stream waters this is HCO_3. If a complexing agent is not present, UO_2^{2+} outside its stability range is hydrolysed to $UO_2(OH)_2$ and precipitated or sorbed from solution. Organic matter under more acid conditions can also increase the mobility of uranium (as a uranyl fulvate complex-Halbach *et al.*, 1980), but the usual role that organic matter plays is to remove uranium from solution. Adsorption onto iron oxyhydroxides is particularly effective between pH 5 and pH 8, but in the presence of the complexing agents discussed above, may be completely inhibited (Rose and Wright, 1980). The adsorption of uranium by clay minerals is well known, the most recent study being by Borovec (1981) who showed that illite and montmorillonite are both effective adsorbers of uranium.

Despite the complexity of the interacting processes controlling uranium in solution, surface water sampling has proved very successful in the search for uranium deposits in the Old Red Sandstone (Devonian)of Caithness (Gallagher *et al.*, 1971; Gallagher, 1972; Michie et al., 1973) and Orkney (Michie and Cooper, 1979). In both these, areas the waters are generally bicarbonate-rich as also they are in south-east Devon.

Sampling

Two hundred and seventy-two samples were collected from small streams in south-east Devon at points of easy access. Thus, the distribution of the samples is not random, but it is irregular. Generally 150ml samples were obtained in polypropylene bottles, but for eight sites in different lithological environments, 21 samples were collected to test for matrix effect. The number of samples collected in a day varied from ten to thirty, and the samples were usually analysed within twenty-four hours of collection and all were analysed within forty-eight hours. The survey period was July/August, 1982. Sample collection presented some difficulties: sites were always chosen upstream of the access points and areas of obvious contamination (such as litter, broken banks, built-up banks, etc.) were avoided. However, some samples that were taken at sites in relatively built-up areas or near farms may have been contaminated. Interestingly, though, no samples taken in these conditions showed extraordinary values when compared with the results from sample points nearby which were free of doubt. Two samples (one from near Ide and another from near Aylesbeare) were so heavily charged with suspended matter that filtering was necessary before analysis, yet they produced the same level of uranium in solution as clean samples from those areas.

Throughout the period of the survey, samples were also repeatedly taken from streams at Whimple, Clyst St. George and near Kennford in order to monitor any effect changing weather might have had on uranium levels. The effect of rainfall was considered likely to be particularly important and it was shown that a Predetermined practice of not obtaining survey samples on days of heavy rainfall or for two days thereafter adequately avoided spurious results.

Analysis

The analysis of the water samples for uranium was carried out using a Scintrex UA3 uranium analyser which works

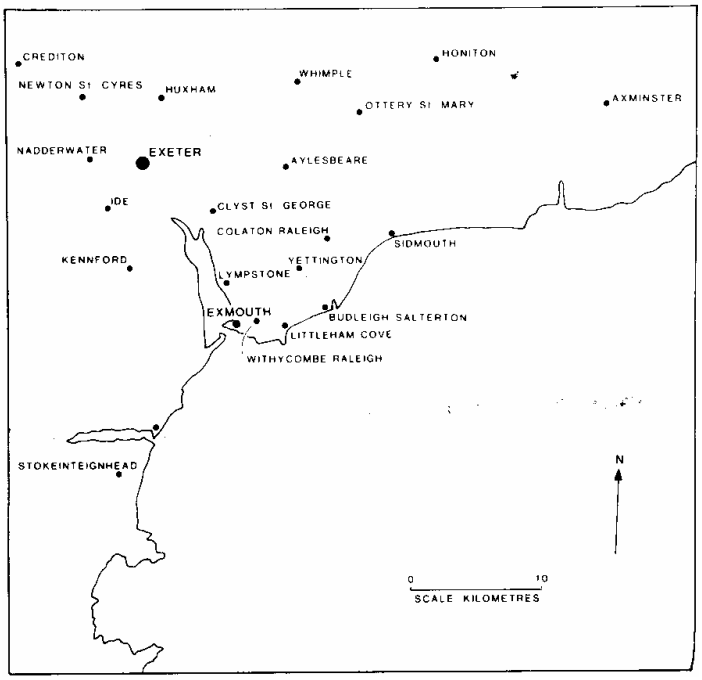

Figure 2. Map of south-east Devon showing the positions of the localities mentioned in the text.

on the principle of laser-induced fluorescence, as described by Robbins (1978). Essentially the method involved the measurement of the characteristic green fluorescence emitted by uranyl compounds under ultraviolet excitation. In the UA3 the excitation is provided by a nitrogen laser which gives a 4ns pulse at a wavelength of 337nm. The fluorescence spectrum produced by the UA3 in a natural water sample shows that, apart from any emission produced by uranium, there is also an emission band due to the presence of dissolved organic carbon. The fluorescence due to organic carbon is, however, very short-lived (4 - 10ns), but that due to uranium persists for much longer. In practice, the fluorescent activity is measured between 1μs and 3 μs after the laser pulse. The instrument is designed to give merely a reading of the activity on an arbitrary scale which can be varied to change the range and sensitivity of the results obtained. As uranium may be present in solution in a variety of chemical forms, each sample is treated by adding 0.8ml of Fluran (sodium pyrophosphate/sodium monophosphate)--a complexing agent which converts all the uranium present to uranyl pyrophosphate.

Because the analyser is calibrated using standard solutions based upon distilled-deionised water, the question arose as to whether or not the natural stream waters from south-east Devon behave in a similar way to distilled-deionised water in their matrix effects. Natural waters carry many elements in solution, some of which may interfere with the fluorescence of the uranium in the sample if present in sufficient concentrations. Among the most important interfering agents are Fe3+ and humic acids, which act as "inner filters"-- absorbing part of the laser pulse, while manganese, calcium and magnesium act as "quenching filters" -- which reduce the lifespan of the uranyl fluorescence. To check on this effect, samples were collected from streams at eight representative sites m different parts of south-east Devon: from Permian breccias at Stokeinteignhead, Permian breccias near Kennford,

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Carboniferous shales and sandstones at Nadderwater, Permian mudstones at Whimple, Permian mudstones at Lympstone, Triassic Pebble Beds at Yettington, Triassic sandstones at Colaton Raleigh and Jurassic shales and limestones near Axminster. Each sample was then used as the matrix for a series of standard uranium solutions ranging from lppb to 20ppb, and analysed in comparison to distilled-deionised water standards. Differences between each sample were found to be negligible and the difference in results obtained in comparison with the distilled-deionised water standards over this range was found to be less than 5%. No correction for matrix effect has therefore been applied to any of the results obtained in this survey, and in consequence the values (quoted fully by Durrance, 1983) are only correct to 0.1ppb.

Results

The distribution of sampled points with the results there from subdivided into representative classes, are given in Figure 3. This clearly shows that the main area of high uranium concentrations in stream water occurs in a northward trending band extending from Littleham Cove on the coast to near Aylesbeare inland. The coincidence of sites where the uranium level is greater than 5ppb with the area underlain by the Littleham Mudstones is noteworthy. The westward extent of the Littleham Mudstones in Figure 1 is taken from Henson (1971) in the southern part of the map, and the most easterly occurrence of the sandstone units within the mudstones of the Aylesbeare Group (Lower Marls with occasional sandstone) shown on the Geological Survey 1:63360 Exeter Sheet (325), in the central and northern part of the map.

A second area where values exceed 5ppb is found around Kennford, where the underlying lithology is breccia. Values between 3ppb and 5ppb also occur in the breccia area west of Dawlish, and an isolated value greater than 5ppb occurs with the breccias in the Crediton trough. Thus although there does appear to be a zone of high values associated with the breccias this is not very consistent and South of the Teign Estuary no value exceeds 3ppb.

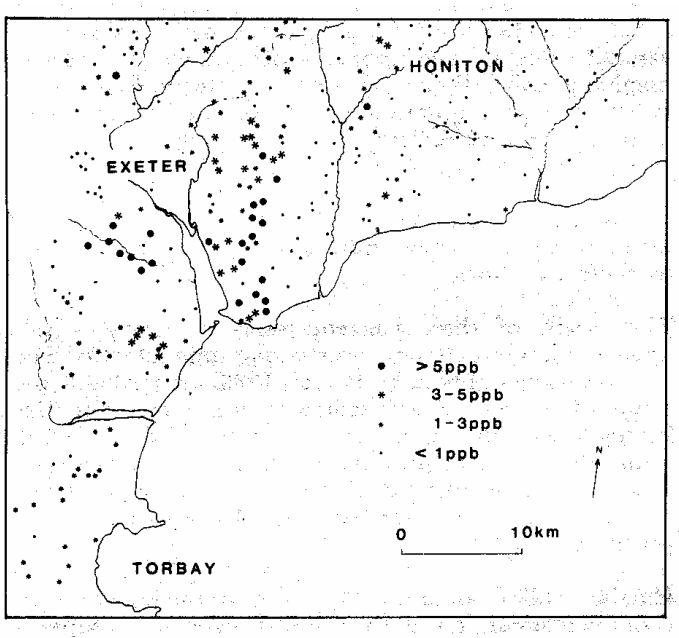

Figure 3. **Results of stream water uranium survey collected into major classes.**

The third area where values are reasonably high is in the belt of ground trending north from the coast at Sidmouth. Here only a single value greater than 5ppb occurs, but there are several other well-separated sites with values between 3ppb and 5ppb. Together these sites show a coincidence with the area underlain by the lower part of the Mercia Mudstone Group.

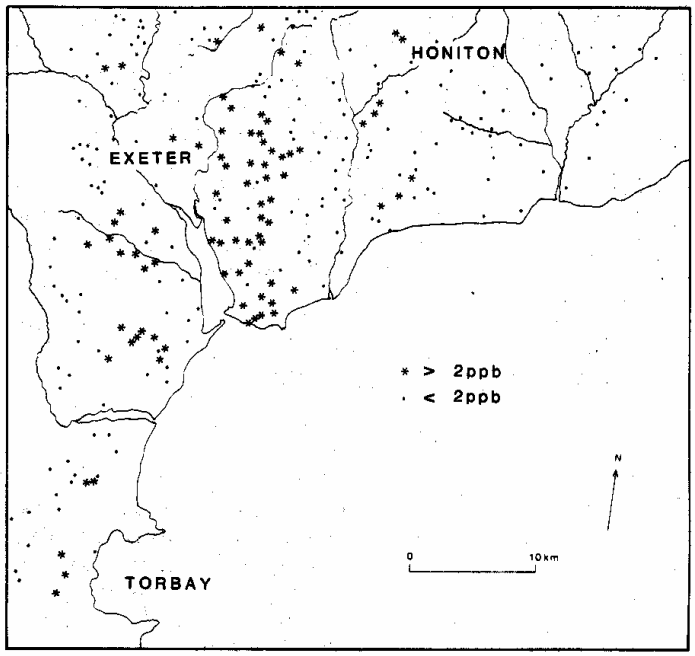

Figure 4. **Distinction between normal and high (greater than 2 parts per billion) sites of stream water uranium.**

Apart from the displacement of high values in the downstream directions there are no other areas where the uranium level can be considered to be particularly high, but it should be noted that not all the. westerly displacement from the Littleham Mudstones is due to stream flow. The Stream which rises east of Littleham Village and flows to the sea at Exmouth has been sampled at five separate points. Those sites on the Littleham Mudstones produced values of 6.Sppb and 7.Sppb, but down-stream the sites on the Exmouth Sandstone and Mudstone Formation show values of S.3ppb, 3.1ppb and finally 2.7ppb over a distance of only 3 km to the sea. If this decrease in value is taken to be the result of loss of uranium by precipitation and adsorption, plus the effect of dilution caused by the addition of "fresh" water, then the maintenance of high values in the westerly flow of other streams and, indeed, the sporadic occurrence of a slightly high value downstream, probably indicates the presence of minor sources of uranium within the Exmouth Sandstone and Mudstone Formation.

Treating values greater than 2ppb as significantly higher than background (lppb and less) gives the data distribution shown in Figure 4. This map clarifies the

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importance of the Exmouth Sandstone and Mudstone and Littleham Mudstone Formations, in controlling the distribution of high values east of Exmouth, and of the lower part of the Mercia Mudstones in the ground between Sidmouth and Honiton. It also shows that south and west of Exeter, the occurrence of high values definitely appears to be related to the breccia formations.

Mineralogical Associations

N. L. Jefferies (unpublished B.Sc. thesis, University of Bristol) has studied the uranium-bearing nodules from the Littleham Mudstones using CR 39 nuclear track plastic. In this technique, radiation damage suffered in the plastic by the passage of alpha particles when a sheet of plastic is placed in contact with a cut surface across a nodule, is made visible by placing the plastic, after a suitable period of exposure, in NaOH. The NaOH causes preferential etching along the radiation damaged pathways, and by re-positioning the etched plastic over the nodule the distribution of the alpha-emitters in the nodule is shown. He found that there are two ways in which uranium minerals occur within the nodules.

(1) In unoxidised form: which represent the primary occurrence of uranium sites.

(2) In oxidised form: which represent secondary mobilisation and precipitation of uranium.

Jefferies showed that both these assemblages occur in close proximity to each other. This unusual close association of primary and secondary minerals is probably caused by the presence of high concentrations of vanadium within the nodules; secondary oxidised uranyl ions would be precipitated as insoluble vanadates close to the sites of solution of the primary uranium. Thin veins of secondary uranium minerals may also penetrate the pale green haloes away from the nodules, but this is uncommon.

Primary uranium minerals occur at the boundaries between light-coloured and dark-coloured rings within the nodules and along shrinkage cracks within the nodules, but are usually in greatest concentration in the cores of nodules. Opaque minerals also line shrinkage cracks as well as occurring within the body of the nodules. The mode of occurrence of the primary uranium is as either colloidal films around grains of other minerals or as amorphous patches, depending upon the size and nature of the grains in the host material. Opaque minerals particularly form sites for the occurrence of colloidal films.

As a result of these observations Jefferies concluded that the main mineralisation occurred after the growth of the nodules was complete. Possibly the nodules formed earlier by deposition of some metals (particularly vanadium) at sites occupied by organic debris within the sedimentary rocks. However, the characteristic Leisegang diffusion rings which occur within the nodules could have formed at any stage in their development. Jefferies suggested that the main mineralisation may be as young as the Tertiary.

A large nodule, similar to those occurring in the Littleham Mudstone Formation , has been found in the lower part of the Mercia Mudstone Group at Sidmouth by Mrs. Edmonston of Ottery St. Mary. In all important respects this nodule is like those from lower in the New Red Sandstone sequence, but the host lithology is rather more arenaceous than for those which occur in the mudstones at Littleham Cove.

Concentrations of Rare Elements and Hematisation of the New Red Sandstone

A problem associated with the mechanism by which hematisation of the red-beds was inhibited in the vicinity of the concretionary nodules in the Littleham Mudstone Formation, suggested by Durrance et al., (1978) was raised by Sellwood (1979), and concerns the need to maintain organic matter as precipitation nuclei within the sedimentary host for a long period of time (perhaps as long as 35Ma). Jefferies' results, taken in conjunction with findings of Harrison (1975), however, suggest that the answer to this problem may lie in early centres of heavy metal precipitation maintaining reducing environments.

One of the most obvious features of the nodules, though a point that is rarely stated, is that not all the nodules are uranium-bearing; yet all show the characteristic pale green halo. In general the nodules typically contain vanadium, nickel, copper, cobalt, arsenic and silver in addition to any uranium. Minerals that are found include native copper, native silver, bornite, chalcocite, chalcopyrite, covelline, maucherite, niccolite, rammelsbergite, vanadian mica, malachite and freirinite, plus possibly modderite. Evidence for the sequence of formation of these minerals is fragmentary, but nickel arsenide (particularly maucherite) was found by Harrison (1975) to line veinlets of native copper which it thus probably preceded. In turn, niccolite and rammelsbergite probably preceded the native metals, although the nickel arsenide contains a small amount of copper. The coffinite shows some vanadium content but no copper, nickel or cobalt, and Harrison proposed a paragenetic link with the vanadian mica. The sporadic occurrence of copper-dominant concretions appears to be unconnected with either the vanadium-dominant or nickel arsenide-dominant concretions, suggesting random precipitation events with no constant paragenesis. Harrison thus proposed an overall cogenesis of vanadian mica, sulpharsenides, coffinite and native metals. Jefferies' work, however, clearly indicated a late origin for the uranium-rich films.

The origin of these concentrations of heavy metals appears to have extended over a long time interval. The finned varieties of nodule (Perutz, 1939) are probably the result of growth in soft-sediment conditions, the fins having formed in response to hydrostatic pressure causing radial displacement into lower pressure, weak pathways in the enclosing sediment (Harrison, 1975). In contrast, the more spherical concretions grew by post-compaction processes.

Harrison (!975) noted that the source of the heavy metals is problematical, in that there is little evidence to Support their derivation from the red host rocks. Indeed, the concretionary enrichment in copper, arsenic, cobalt, nickel,

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chromium, arsenic, sulphur, vanadium and uranium, on the contrary, supports the presence of an exogenous source for their solutions. Harrison favoured the concept of deep-seated hydrothermal or hot spring sources as Hawkes (1974) had proposed for the tin metallogenesis of south-west England. He related the activity to deep crustal tensions arising from plate movements and the opening of the North Atlantic, and therefore implied the importance of fracture permeability in the system.

The spatial association between the uranium mineralisation and the heavy metals which preceded it, however, clearly shows that the concretionary sites remained areas of reducing conditions for a long time. Clearly, uranium precipitation and iron mobility form an environmental association, while uranium mobility and iron precipitation are a contrasting environmental association. The early--late age of formation of the concretionary nodules within the compaction history of their host sediments, thus can be seen as providing local reducing environments, which could have persisted until recent exposure by erosion allowed oxidising conditions to develop. Not only would the heavy metals have acted as precipitation sites for later uranium in solution passing through the sedimentary sequence, but could also have acted as the localised inhibitors of haematisation.

Interpretation

Some 20Ma after the end of the Variscan orogeny, renewed magmatism occurred in south-west England. This was probably related to thermal activity in the mantle and, in turn, gave rise to hydrothermal circulation and mineralisation during New Red Sandstone times.The main phase of metalliferous mineralisation in Cornubia occurred about 270Ma ago.

Apart from manganese deposits between Huxham and Newton St. Cyres, direct evidence of mineralisation occurring within the New Red Sandstone sedimentary rocks of south-east Devon is first encountered in the Littleham Mudstone Formation. Here the results of the stream water uranium survey suggest the presence of uranium-bearing nodules over a wide area. Although the age of the Littleham Mudstone Formation is not known, its position fairly high in the Permian succession indicates that any correlation between the earliest episodes of hydrothermal circulation proposed by Harrison (1975) to account for the formation of the nodules, and the main phase of metalliferous mineralisation in Cornubia, is unlikely. Moreover, as similar mineralisation occurs in host rocks of Triassic age, it would seem logical to place these events as commencing at the earliest in the Triassic, possibly correlating with 210Ma mineralisation event in west Devon and Cornwall.

Dispersed mineralisation through the mudstones of the New Red Sandstone requires the pervasive movement of groundwater as well as the presence of sites at which precipitation could occur. Groundwater movement over large distances through low permeability rocks by intergranular flow is, however, not possible. A more realistic approach is to consider that groundwater movement is essentially restricted to major fracture zones in otherwise impermeable rocks, but that charging of drained aquifers can occur if these are encountered. Within the New Red Sandstone sequence in south-east Devon the main aquifers are the Dawlish-- Clyst Sandstones and the Sherwood Sandstone Group (Budleigh Salterton Pebble Beds and Otter Sandstone Formation). Fracture zones carrying upward moving groundwater, with the heavy metals in solution, could have charged these horizons, but subsequent drainage through the adjacent rocks would have occurred. The main drainage direction appears to have been down from the Sherwood Sandstone Group into the Littleham Mudstones, although the true picture is confused because movement will only have been recorded where suitable precipitation sites were present. Certainly upward drainage into the Mercia Mudstones also seems to have occurred, where the mineralisation appears to be much more dispersed--possibly reflecting the more arenaceous character of this unit. The sporadic occurrence of concentrations in the Exmouth Sandstone and Mudstone Formation may indicate either limited charging of the sandstone aquifers within the formation or upward drainage from the Clyst Sandstones. Occasional precipitation sites for the heavy metals appear to have been present in the breccias underlying the Dawlish--Clyst Sandstones, but whether or not this aquifer became charged to the same extent as the Sherwood Sandstone Group is difficult to say. Possibly the limited lateral development of these aeolian sands meant that they were not the open system needed to permit extensive charging to take place. Some charging of the arenaceous megafacies of the Mercia Mudstones, and subsequent drainage, also appears to have occurred to account for the widespread development of the "fish eyes".

The precipitation sites for the heavy metals appear to have been organic debris which was preserved in the mudstone formations, but some precipitation at sites of permeability contrast, as suggested by Harrison (1975), could also have occurred. The early phases of mineralisation do not. appear to have been uranium-rich, the uranium was introduced at a later stage. Post-New Red Sandstone mineralisation has probably occurred, and Durrance and Heath (1985) show evidence that it is still taking place.

The lateral limit of charging of an aquifer by groundwater movement from a fracture zone is about five times the depth extent of the fracture zone (Goyal and Kassoy, 1977; Goyal, 1978). Thus if the occurrence of the nodules at Littleham Cove is taken as marking the position of the fracture zone-- a concept supported by the modern zone of rising groundwater passing through this area (Durrance and Heath, 1985) and if precipitation sites are uniformly distributed throughout the Littleham Mudstone Formation, then a depth extent for the fracture zone of about 3 km is suggested. This value is derived simply by noting the distance over which the nodules occur to the north, as shown by the results of the stream water uranium survey. However, other fracture zones that are present, particularly an easterly extension of the fault that bounds the southern margin of the Crediton trough (which is coincident with another modern belt of rising groundwater shown by Durrance and Heath, 1985), may also have allowed the aquifer to be charged along lines further north. In fact, this seems necessary to account for the northward extent of the mineralisation in the lower part of the Mercia Mudstones

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(as shown by the results of the stream water uranium survey), and to produce the manganese mineralisation between Newton St. Cyres and Huxham. The implication, if this is correct, is that the precipitation sites are not uniformly distributed through the Littleham Mudstones. Perhaps the occurrence of the main nodule zone east of Exmouth (again as shown by the results of the stream water survey) is due to the spatial coincidence of numerous precipitation sites in a mudstone formation, the position of these beneath the major aquifer of the region, and the presence of a fracture zone which permitted the ascent of thermal waters to charge the aquifer.

Regarding the origin of the solutions, it is worth noting that the source of the heavy metals does not lie within the New Red Sandstone sedimentary rocks (Harrison, 1975). Also, that although Ball et al., (1979) looked outside the sedimentary rocks to the granites for the source of uranium mineralisation in west Devon and Cornwall, the bulk of the cross-course mineralisation originated by leaching of the Devonian and Carboniferous sedimentary rocks (Bull, 1982). It is therefore interesting to find that much of the chemistry of the cross-courses is also represented in the mineralization of the New Red Sandstone. A diagrammatic illustration of the hydrothermal system that could have produced the New Red Sandstone mineralisation is shown in Figure 5.

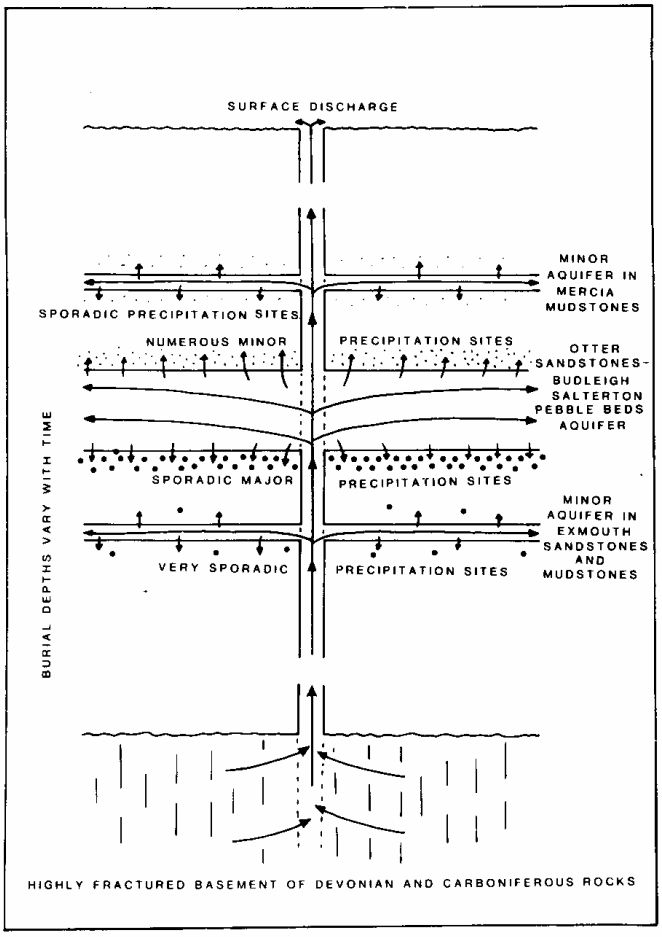

Figure 5. Model of groundwater movement up deep seated faults, charging of Permian-Triassic aquifers and formation of nodular mineralisation.

Implications

The hematisation of the New Red Sandstone appears to have occurred after the first phase of metalliferous mineralisation, and was very pervasive. In contrast to the introduction of metals from outside the sequence, the hematisation appears to have been caused by the redistribution of iron within the various lithologies, and this has occurred on a vast scale. The iron content of the pale green mudstones within the New Red Sandstone, even that of the black mudstones near the top of the succession, is typically about 2% less than that of the hematised mudstones. The iron content of the pale green haloes around the concretionary nodules is also 2 - 3% less than that of the hematised mudstones. Solution of ferrous iron, its mobilisation and precipitation after oxidation as ferric oxyhydroxide by in situ alteration is only likely to be effective where the permeability of the rocks is high and where groundwaters have a high Eh. Some addition of uranium may also have occurred by transport in waters with a high Eh, and the two processes may be linked.

Whereas precipitation of ferric oxyhydroxide requires an increase in pH with little change in Eh, uranium precipitation will occur if there is a decrease in the Eh of the environment. The supply of groundwater to the mudstones must, however, again be related to charging--discharging aquifers within the New Red Sandstone sequence. As suggested by Durrance *et al.,* (1978), the large-scale involvement of groundwater in this process may only have become possible for the first time when the climate of the area changed to humid conditions near the start of Jurassic times. The source of both the additional iron, and possibly some of the uranium, is thus seen as the aquifer horizons themselves, with in situ weathering of detrital minerals within the succession occurring in the manner suggested by Walker (1967). With the hematisation post-dating any early formation of metal concentrations and occurring at the end of Triassic times, further evidence which supports a correlation between the earliest mineralisation and the 210Ma event recorded in west Devon and Cornwall is afforded.

Finally, if the hematisation process took place under an overburden thickness of about 1000m as suggested by Durrance et al., (1978), it is quite likely that some element of hydrothermal circulation was involved in the movement of the groundwaters which caused the weathering of the detrital minerals in the aquifers.

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Read at the Annual Conference of the Ussher Society, January 1984

A gravity survey of the Polyphant Ultrabasic Complex, East Cornwall.

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E. M. DURRANCE
R. J. JADY

Chandler, P., Davey, R. F., Durrance, E. M. and Jady, R.J. 1984. A gravity survey of the Polyphant Ultrabasic Complex, East Cornwall. *Proceedings of the Ussher Society 6*, 116-120.

The Polyphant Ultrabasic Complex in East Cornwall mainly consists of peridotite, with subsidiary amounts of gabbro and dolerite. Zones of alteration associated with flat-lying shear planes within the peridotite result in the formation of serpentine, chlorite and carbonate minerals. Interpretation of the results of a detailed gravity survey to determine the form of the Complex was rendered difficult because of its presence in the steep gravity gradient caused by the Bodmin Moor granite. However, processing of the results by a two-dimensional interpolation and computer contouring procedure allowed objective production of regional and residual Bouguer anomaly maps. These show that the Complex has little effect on the gravity field and is probably only about 32m thick. It has the form of a fault-bounded slice. The occurrence of other small bodies of peridotite in central south-west England is taken, together with Polyphant, as indicating the presence of a tectonically dismembered ophiolite--flysch complex in the area.

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Introduction

The Polyphant Ultrabasic Complex in East Cornwall lies about 5 km northeast of the Bodmin Moor granite and some 9 km west of Launceston. Although the Complex was originally identified as a picrite by Reid *et al.* (1911) and was, therefore, a somewhat unusual feature in the geology of south-west England, very little modern work has been carried out on it. However, the revision of the Institute of Geological Sciences Sheets 337 and 338 by a team working under contract to the Institute at the Department of Geology, University of Exeter, has allowed a recent detailed examination of the geology of the Polyphant area.

As part of the contract work Stewart (1981) produced a revised geological map of the Devonian and Carboniferous rocks which surround the Complex. The main features of this are shown in Figure 1. The ultrabasic body is poorly exposed, but its outcrop can be seen to extend northwest - southeast for a distance of approximately 2 km, with a maximum width of outcrop of about 0.5 km. It is fault-bounded on all sides.

During the Variscan deformation the Devonian and Carboniferous rocks of the Polyphant area were involved in thin-skinned thrust and nappe tectonics, which disrupted and translated the original stratigraphy northwards for several kilometres (Isaac et al., 1983). Later (post-nappe) Variscan structures are dominated by northeasterly dipping, low-angle extensional faults with associated folds and kink bands. It has been suggested by Isaac et al. (1982) that these late normal faults could be related to the rising to a high level of the Bodmin Moor granite. The long and intense deformation history of the area has resulted in a tectonic stratigraphy consisting of intercalated sheets of widely differing lithologies. Thus, on its northeastern flank the Complex is overlain by Upper Devonian and Lower Carboniferous slates which are interleaved with Upper Devonian shallow-water limestones. The southwestern boundary of the Complex is formed by a high angle fault which dips steeply to the northeast, but this fault may well shallow out at depth.

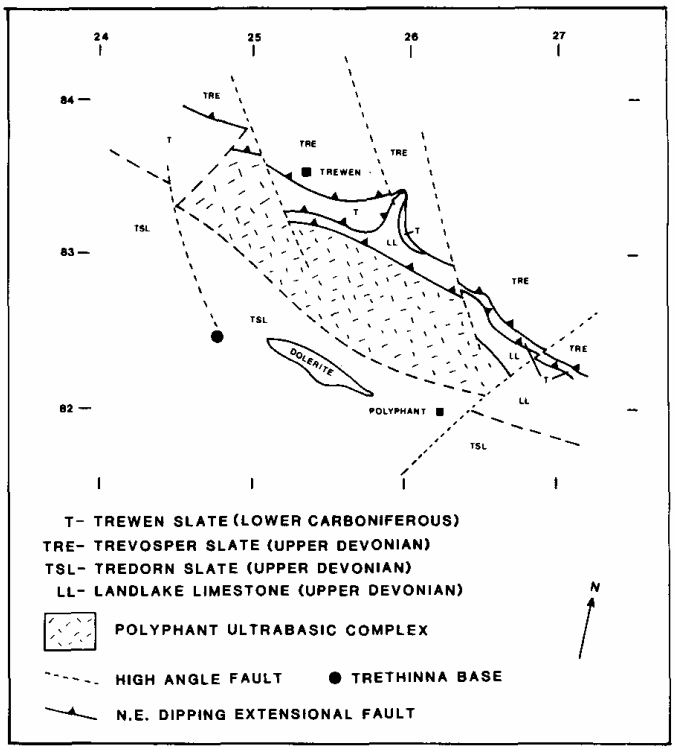

Figure 1. Geological map of the Polyphant area. (Coordinates on Fig. 1-4 are National Grid references for 100 km square SX)

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The Polyphant area also lies within a "wrench" fault zone trending north-northwest -south-southeast which was thought by Dearman (1963) to be of Tertiary age. Stewart (1981), however, suggested that within this zone vertical movements are much more important than strike-slip movements. Turner (I981) has also demonstrated that similar fault zones elsewhere in central south-west England have a history of movement dating back to the Variscan deformation. As can be seen from Figure 1, minor faults which trend between northwest - southeast and north-northwest - south-southeast are abundant in this area and offset the boundary of the Complex.

Lithologically the Polyphant Ultrabasic Complex is dominated by peridotite, but this occurs in association With minor amounts of gabbro and dolerite. The peridotite is variably altered from "fresh" material containing serpentinised olivine, clinopyroxene, brown hornblende and biotite to an assemblage of serpentine, chlorite and carbonate minerals. Because of the poor exposure, the extent of the alteration throughout the Complex is not known, although sparse borehole material and outcrops in the New Quarry (Grid Reference SX 25958258) indicate that the most pervasive alteration is associated with fiat-lying shear zones that cut the body. The borehole material also suggests an interbanding of the peridotite with gabbro and dolerite, but this is never seen in outcrop. It is probably that this interbanding is of a tectonic rather than an igneous origin because of-the apparently small size of the body. However, if the Polyphant Ultrabasic Complex is part of a deep-rooted intrusion, or can be shown to be part of an originally much larger intrusion, then the development of igneous layering cannot be ruled out.

Thus before any ideas about the significance of the Polyphant Ultrabasic Complex could be formulated, its geometry and thickness needed to be known. For this reason it was decided to carry out a detailed gravity survey over the area.

Gravity Survey and Rock Densities

The nature of the regional gravity field in the Polyphant area is well known from the work of Bott *et al.* (1958) and the Institute of Geological Sciences (1975). Polyphant lies within the very steep Bouguer anomaly gradient on the northeastern flank of the negative anomaly caused by the Bodmin Moor granite. Over a northeast - southwest distance of 5 km centred on Polyphant, Bouguer anomaly values decrease by about 20mgal. Neither of the regional gravity maps shows any special feature in the Polyphant area. However, as Bott et al. (1958) established only a small number of gravity stations between Launceston and Bodmin Moor and the Institute of Geological Sciences map was constructed using an average gravity station density of less than one per square kilometre, the absence of such a feature may merely reflect the distribution of the gravity stations.

Because of the complicated geological and geophysical setting of the Polyphant Complex, it was recognised at an early stage in this work that three areas of difficulty could cause problems in the interpretation of the data acquired from a detailed gravity survey of the body. These are:

(1) the steep northeast - southwest gravity gradient produced by the Bodmin Moor granite would make resolution of any small local anomaly at Polyphant difficult to obtain.
(2) the zones of alteration within the peridotite could have produced extensive lateral and vertical variations in the density of the rocks comprising the body.
(3) the steep topographic features which are present in the area, particularly on the sides of the valley of the River Inny, would necessitate the application of large terrain corrections.

Attempts to minimise the effects of these problems were therefore introduced during the conduct of the gravity survey itself. Thus, two scales of survey were employed. To obtain good information on the regional gravity gradient in the area, 53 gravity stations were established in the ground surrounding Polyphant, over an area of 34 krn2, with an average separation of about 1 km. Over the Complex itself, however, the gravity survey was conducted in a series of northeast - southwest traverses about 300 m apart, along which the gravity stations were positioned at approximately 100 m intervals. Cross-linking traverses were also carried out with the gravity stations at an interval of about 200 m. A total of ll9 gravity stations were established over the Complex or in its immediate vicinity.

Positioning of the gravity stations was achieved either by the use of 1:10,000 scale Ordnance Survey maps or derived as part of the topographic survey carried out to obtain the elevation of each station and the details needed for terrain corrections. Surveying was by means of theodolite and subtense bar, supplemented where necessary by surveyor's staff and level.

During the survey work a local gravity base station was established at the road junction at Trethinna (SX 24828253), and marked, by a painted spot on the road surface. This was reoccupied on several occasions during each day's survey work, the elapsed time between successive base station readings always being less than two hours. Tidal and instrument drift variations between these readings were assumed to be linear. The results given in Figure 2 are arbitrary Bouguer anomalies related to this local base station value of 4736.64mgal. The Trethinna base was tied in with the national gravity network via the gravity station at the Liskeard Fundamental Bench Mark (SX 22246355) which has a value of 981081.076gals. Three alternate repeat readings were taken at the Liskeard and Trethinna base stations, with the elapsed time between successive readings at each station always being less than 75 minutes. From this work the difference between the Trethinna base and the Liskeard gravity base station has been established as 2.15mgals (after correction for latitude and elevation differences). However, because the interpretation of the gravity data around Polyphant is only concerned with the resolution of a local feature within the main gravity field, it has not been considered necessary to recompute the Bouguer anomaly values.

The survey instrument used throughout this work was a LaCoste and Romberg Model G gravity meter, and the

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local Bouguer anomalies are considered to have an accuracy of ±0.01mgal. The error in connecting the local anomalies to the national network is, however, about ±0.02mgal.

Rock densities were obtained from the Devonian slates that surround the Polyphant Complex, and from both "fresh" and altered varieties of the peridotite. Exposures of unweathered material are absent in the area and so the derived densities may be somewhat lower than their true *in situ* values. Water-saturated densities were measured by weighing in air and water. The mean value obtained for Devonian slate is 2500kgm^{-3} (standard error of the mean ±100), while that for "fresh" peridotite is 2880kgm^{-5} (standard error of the mean ±40) and that for altered peridotite is 2870kgm^{-3} (standard error of the mean ±20). These values clearly show that a distinct density contrast of 380kgm^{-3} occurs between the "fresh" peridotite and the Devonian slates, and the highly altered peridotite still has a distinct density contrast of 370kgm^{-3}.

Production of the Bouguer anomaly maps

The first step in the production of a Bouguer anomaly map is the use of an interpolation scheme that will convert, to an acceptable degree of accuracy, gravity measurements (corrected for latitude, height, and instrument drift), taken at irregularly spaced intervals, to values at the nodes of a regular rectangular grid. These grid values can then be used in a contouring routine to produce Bouguer anomaly maps, from which direct interpretation of the nature and extent of any local anomaly can be made. They are also then available in a suitable form for further analysis such as a two-dimensional Fourier representation, or downward continuation.

Several schemes exist for interpolating a function in two dimensions given at irregularly spaced points. For example the Lagrangian interpolation method, which is based on approximating the given function by polynomials, or cubic splines, is commonly used. Also widely used is the Aitken successive linear interpolation method. This method, which does not explicitly calculate ' the interpolating function, is most accurate when there is a degree of symmetry in the distribution of the data points about the grid point at which the interpolation is being made. Some accuracy is lost, however, when values outside the domain of the observational data are extrapolated, and spurious results can be obtained.

The method finally used here is similar to the Aitken scheme, and is based on the work of Falconer (1971). It is superior to the other schemes in that it yields data in a form more suitable for later contouring procedures. In this, the interpolated value at a grid point is determined by a least squares fit of a paraboloidal surface to the suitably weighted surrounding observations. Only those observations that lie within a circle of radius R centred on the grid point are used, and the weighting factor W, applied to such an observation at a distance D from the grid point is given by $W= ((R-D)/D)^2$, where R>D>O.

Four parameters are needed to determine the interpolating surface, so that it is necessary to ensure that the radius of the circle is large enough to include at least four points for each interpolation, and preferably six. Including a greater number of points increases the radius of the circle to cover a greater proportion of the map, and can lead to over-smoothing and increase the computing time.

The computer contouring procedure of Heap and Pink (1969) relies implicitly on the regular grid pattern of values generated by the interpolation process. For the outcome to be successful, care must be taken to ensure that a sufficiently fine grid, consistent with the data and the desired detail in the final map, is defined from the outset.

The final numerical scheme was tested with analytical data. The main objectives were to establish the minimum number of data points below which the contouring becomes unreliable, the appropriate grid size to use, and determine how the distribution of points affects the final map; Different grid sizes were tried, and the optimum size was found to be 21 x 21 over the sampled area; increasing the number of mesh points beyond this size did not give

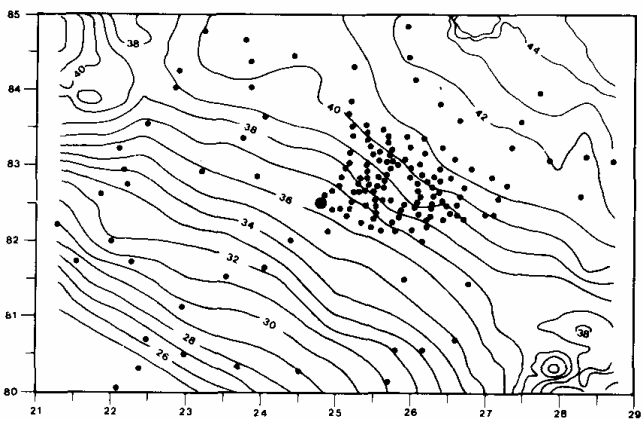

Figure 2. Results of the gravity survey of the Polyphant area. Isogals constructed using all gravity stations. Trethinna base station indicated by larger point.

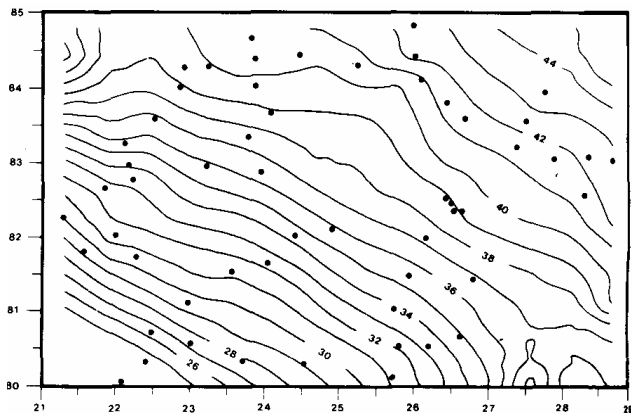

Figure 3. Isogals in the Polyphant area constructed using only those gravity stations that were positioned away from the Ultrabasic Complex.

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better definition.

The final Bouguer anomaly map (Fig. 2), clearly shows the strong regional gradient steadily increasing towards the northeast, with a slight southwest deviation in the contours over the body itself. Also noticeable is the corner effect, in which spurious asymmetry contours are introduced because of the data particularly near the corners. Figure 3 shows the contours obtained without including the gravity measurements taken over the body. As before, the regional gradient shows clearly, and the only difference in this case is that the contours over the position of the Complex curve slightly towards the northeast. Finally, the detailed contour map shown in Figure 4 gives the most accurate and detailed representation of the local anomaly.

The body only produces slight perturbations in the regional pattern of the isogals, and must therefore be of limited extent and depth. As an aid to interpretation, the effect of a positive lens-shaped body on the regional gradient can be seen in Figure 5. The lens-shaped body produces a similar deviation in the isogals.

Discussion

From the Bouguer anomaly maps shown in Figures 2, 3 and 4, it can be seen that the Polyphant Ultrabasic Complex has no significant local gravity anomaly associated with it. Within the accuracy limit of ±0.01mgal of the data, this suggests that the body. has a thickness of only approximately 32 m, assuming a horizontal slab with a uniform density contrast of 370kgm^{-3} based upon the density of altered peridotite. In practice, as areas of "fresh" peridotite occur within the body, the actual thickness could be somewhat less. The evidence from the gravity survey thus indicates the presence of a thin, fault-bounded slice for the form of the Complex. It is certainly not a deep rooted intrusion.

Several other small lenses of similar peridotite occur within the thrust and nappe terrain of central south-west England. For example, a peridotite mass occupying an area approximately 100 m by 50 m occurs southeast of Callington (SX370689) and another body approximately 40 m by 20 m is found near Trekelland (SX345803). These two peridotites and the majority of Lower Carboniferous basaltic pillow lavas and dolerites in central south-west England occur within the Greystone Nappe (Isaac *et al.*, 1982). The pillow lavas and dolerites which occur within this nappe show ocean floor geochemical affinities (Chandler and Isaac, 1982).

In the model of Isaac et al. (1982) for the evolution of the thrust and nappe terrain of central south-west England during the Variscan deformation, the Polyphant Ultrabasic Complex prior to thrusting would have occupied a position immediately below the Greystone Nappe. It is therefore possible that the Polyphant Complex was also temporally as well as spatially related to the pillow lavas and dolerites. Occurring with these basic and ultrabasic rocks are radiolarian cherts, olistostromes and flysch. Lithologies of this type are typical of ophiolite associations. Elsewhere in the world (such as the Tethyan ophiolites of the "croissant ophiolitique") ophiolites are typically tectonically dismembered, and the more general

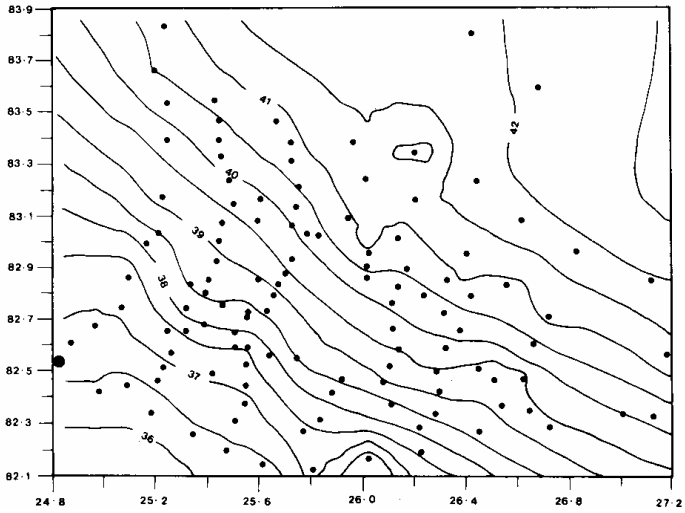

Figure 4. Isogals in the area of the Ultrabasic Complex

term "ophiolite-flysh complex" aptly describes the whole association (Hall, 1980).

Although the original positions of the basic and ultrabasic igneous rocks together with the radiolarian cherts, olistostromes and flysch in central south-west England, is not known, they are part of the same tectonic terrain. The implication that the Polyphant Complex, with the associated ultrabasic rocks at Callington and Trekelland, form part of a similarly dismembered ophiolite--flysch association is therefore clear. The presence of ocean-floor environments within the Devonian and Lower Carboniferous of East Cornwall is consistent with the plate tectonic model of southward subduction for the Variscan of south-west England, originally proposed by Mitchell (1974), and the general model for the plate tectonic setting of ore deposits associated with subduction given by Beckinsale and Mitchell (1981).

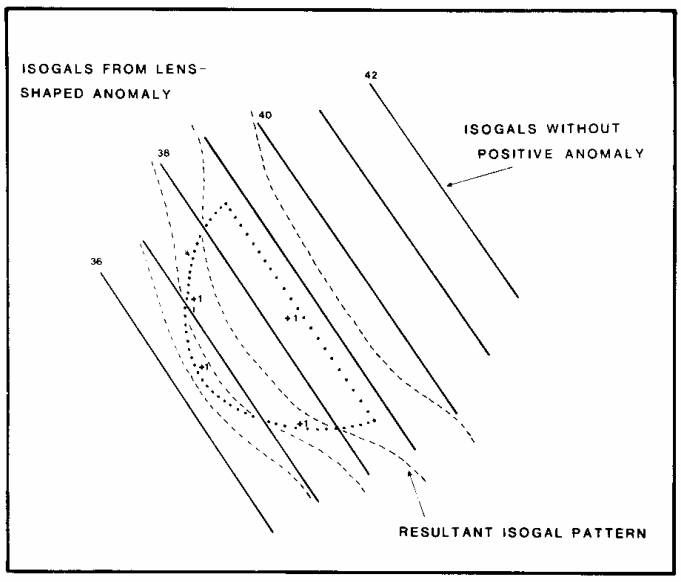

Figure 5. Diagrammatic representation of the effect of a thin lens-shaped ultrabasic body on a linear gravity gradient.

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Read at the Annual Conference of the Ussher Society, January 1984

Garnetiferous phosphatic nodules within the Upper Devonian-Carboniferous Transition Group, near Boscastle, north Cornwall.

D. S. ANDREWS*
G. M. POWER

Andrews, D. S. and Power, G. M., 1984. Garnetiferous phosphatic nodules within the Upper Devonian-Carboniferous Transition Group, ne r Boscastle, north Cornwall. *Proceedings of the Ussher Society, 6*, 121-128.

Rare zoned nodules in graphitic shales are reported from the upper Devonian-Carboniferous Transition Group near Boscastle, north Cornwall. The nodules have cores of apatite + garnet surrounded by a garnet + quartz zone and an outer rim of chlorite and graphite dominated by spessartine-almandine garnets. These garnets have a sectorial preferential growth habit and show strong chemical zoning with Mn enriched margins.

A model for formation of the nodules involves phosphorite being concentrated diagenetically in carbonaceous muds and acting as a nucleus for diagenetic encrustation by iron and manganese minerals, possibly carbonates rather than oxides. Subsequent metamorphism retained the chemical zonation forming apatite rich cores and spessartine-almandine garnet rims.

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Introduction

The rocks exposed around Boscastle, north Cornwall are of Upper Devonian to Carboniferous age. They have suffered extensive deformation and low-grade regional metamorphism during the Variscan orogeny and consist predominantly of slates passing upwards into minor basic volcanic rocks and Carboniferous shales and sandstones.

Freshney *et al.* (1972) review the earlier work on the metamorphism of these rocks and whilst clearly emphasising the caution that must be exercised in an area where the geology is dominated by the juxtaposition of tectonic slices, they describe the metamorphism as low-grade characterised by mineral assemblages in the slates of quartz, albite, muscovite, chlorite and locally chloritoid. They interpret the textures as indicating several periods of metamorphism. Primmer (1983) working in the Tintagel area confirmed the polyphase nature of the metamorphism and the low-grade mineral assemblages but made the important observation that the mineral assemblages in the aluminous slates would be limited by the chemical composition of the slates at the expense of more ferromagnesian minerals such as biotite and garnet.

Records of the occurrence of garnets in slates of the general area do exist. Phillips (1928) claimed that the Delabole Slates which occur south and east of Tintagel frequently contain minute garnets. He analysed a garnet concentrate from a Geological Survey specimen (E6514) from the Delabole Slates, Villapark, near Camelford and found the garnet contained manganese with a minimum of 12.5% of the spessartite molecule. Phillips clearly argued that the presence of manganese would make the garnet stable at a lower grade than that of almandine garnet, the zone mineral for pelitic rocks. Primmer (1982) also found manganiferous garnets, with cores containing up to 28.2% MnO, in graphitic slates associated with the Tintagel Volcanic Formation. Neither of these authors suggest how the garnets were able to concentrate manganese to this extent although Phillips (1928) does mention indications of manganese in Devonian sediments and former manganese workings in Carboniferous rocks.

The important influence of chemical composition on the occurrence of garnet must be emphasised, On the one hand the chemical composition of the rocks may preclude the formations of garnet at its usual grade of metamorphism and on the other hand the availability of manganese may stabilise garnet at a lower grade than usual. Therefore, when one of us (DSA) found several examples of well-preserved lenses containing abundant small garnets within the Transition Group near Boscastle, it seemed clear that they were Of interest and worthy of further investigation as to their nature and origin.

Field Relationships

Following Freshney *et al.* (1972) the rocks of the Boscastle area may be subdivided as shown in Figure 1. The Upper Devonian Tredorn Slates, a monotonous series of coarse grained, grey-green, rust-spotted slates pass upwards into a dark grey slate which spans the Devonian-Carboniferous junction. Lateral equivalents of this unit appear to be very variable and it has been termed the Transition Group. A small slice of the Tintagel Volcanic Group is thrust over the Transition Group and then the Namurian Crackington Formation, here composed of shales and thin sandstones, is thrust over both of them.

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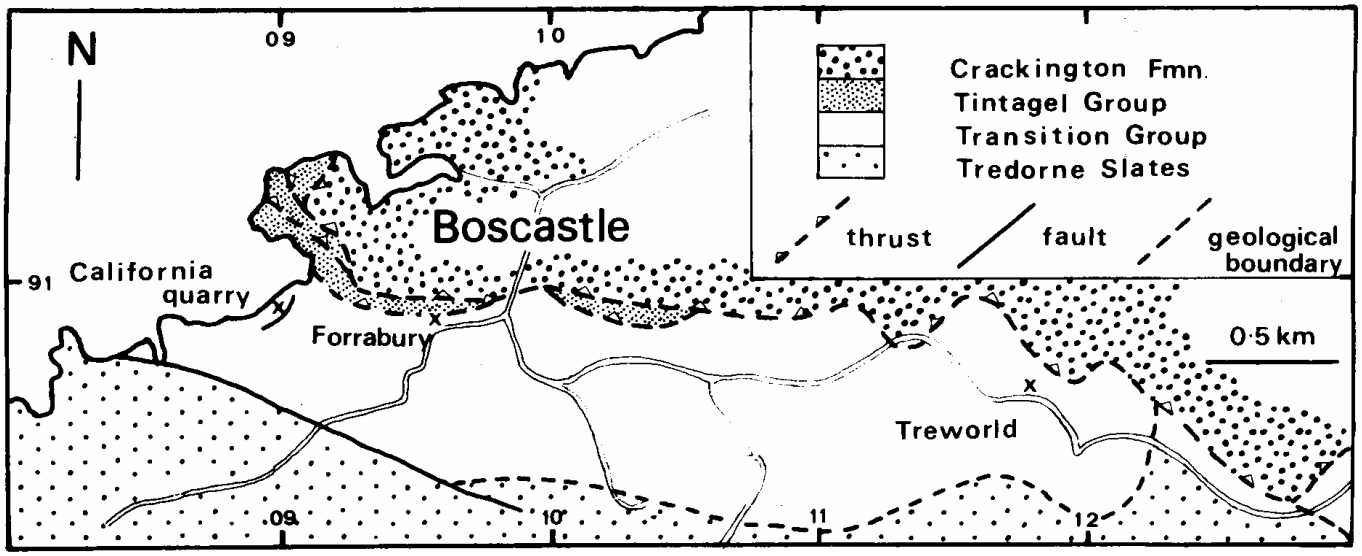

Figure 1. Outline geological map of the Boscastle area. Sample localities marked by a cross.

The principal locality at which garnetiferous lenses were found is in a roadside section (SXl1809056) west of Treworld. Here the Transition Group consists of pale blue-grey slates which become increasingly silicified towards a thrust contact with the overlying Crackington Formation. Four lenses up to 350 mm by 80 mm in size were discovered with their long axes broadly parallel to the cleavage of the slates. The lenses have been disrupted by the deformation which produced minor folding of the cleavage. The lenses exhibit a zoned structure with a very soft quartz-veined grey core surrounded by a 5 mm rim of hard black pelitic material containing light brown euhedral garnets up to I mm in size. The junction between these two zones is sometimes marked by iron-staining.

A second locality occurs along the path to Willapark (SX09619089) east of St. Symphorian's Church, Forrabury where a 200 mm by 20 mm lens was found in dark grey quartz-veined deformed slates. Brown garnets up to 2 mm in size are well-developed on the outer margin of this lens.

Although other outcrops of the Transition Group were searched for further garnetiferous lenses no signs of garnet were found. Samples of dark brown earthy lenses from California Quarry (SX090908) were examined carefully for garnets but none were found. The garnetiferous lenses appear to be rare and are possibly confined to the upper part of the Transition Group.

Zonal structure of the lenses

Thin sections were cut from several of the lenses and they all showed broadly similar features and in particular often have a well-developed zonal structure. However, the zonal structure was very poorly preserved in the specimen from near St. Symphorian's Church. Accordingly, the main description that follows is based on one particular specimen from near Treworld selected as showing the representative features of the lenses.

The main features of the lens are shown in Figure 2, a low magnification photomicrograph of a thin section cut across the centre of the lens. The central zone is composed of garnets typically less than 0.15 mm in diameter in a dark groundmass. Although both garnets and groundmass, contain inclusions of quartz and graphite, even under high magnification no crystal structure may be resolved in this dark groundmass. Quartz veins, possibly of more than one generation stand out prominently in the central zone approximately normal to its margins but showing evidence of some deformation subsequent to their formation. Although some of these veins may be traced into the outer zones they are not nearly so prominent there. Zones composed mainly of quartz and garnet are asymmetrically developed either side of the central zone. The garnets are poorly formed and full of inclusions, although sometimes they have a clear rim and they become larger towards the outer parts of the zones.

The most prominent development of the largest garnets is at the outer margin of the quartz + garnet zones where the garnets grow in the surrounding phyllosilicate + quartz matrix. Often they contain apparently radially arranged inclusion-free rims. The garnets decrease in size away from the lens into the phyllosilicates. Clearly the most favourable conditions for garnet growth occurred at the boundary between the lens and the phyllosilicates where there was the most readily available supply of all the elements necessary for garnet growth. Study of polished thin sections in reflected light shows also that the greatest concentrations of graphite are to be found here.

There is a change in the mineral composition of the phyllosilicate zone with distance away from the edge of the lens. Close to the lens chlorite is dominant whilst further away muscovite and quartz become more important.

Other minerals that may be seen in the phyllosilicate zones include an opaque mineral with the characteristic shape of ilmenite and very rare euhedral prisms of tourmaline.

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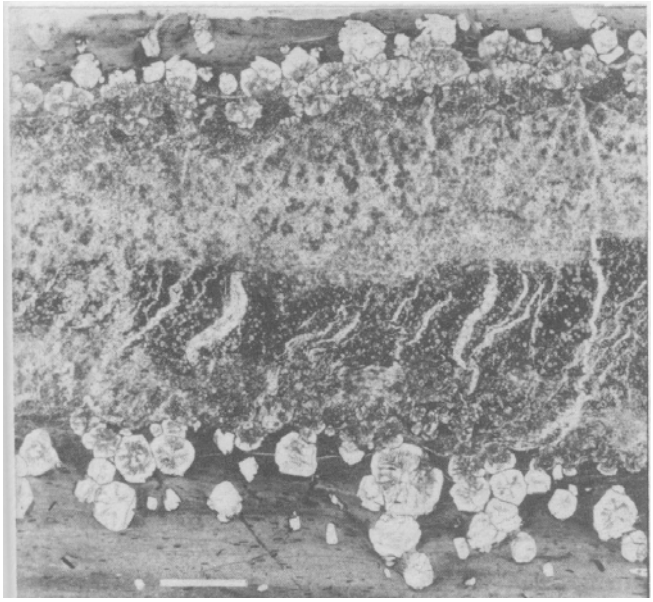

Figure 2. Photomicrograph of zoned nodule from Treworld. Scale bar is 2 mm.

In summary, a well developed zonal structure was discovered in the lenses consisting of a central zone of garnet in a dark unresolved groundmass surrounded by an outer zone of quartz and garnet. At the junction of this outer zone with the phyllosilicates the greatest concentration of graphite occurs and the garnets are most preferentially developed.

Apatite in the central zone of the lenses

The nature of the dark groundmass in the central zone was investigated using scanning electron microscopy. Figure 3 shows a scanning electron microscope photograph of a polished thin-section where the scale bar represents 10 microns. In the lower part of the picture the outline of a euhedral garnet can be observed whilst the dark groundmass is beginning to be resolved into what could be poorly formed hexagonal prisms. Close examination also reveals wisps, some with a spiral structure, possibly of graphite. Energy dispersive X-ray spectra of the hexagonal prisms showed the presence of calcium and phosphorus within them. X-ray diffractometry on a crushed sample from the central zone of a lens proved it to be composed of a mixture of garnet and apatite with minor quartz.

The discovery of apatite within the lenses is considered to be of fundamental importance in the origin of the lenses and will be discussed further in a later section.

Chemical composition and zoning of the garnets

An initial series of chemical analyses of the garnets have been made Using the energy dispersive microprobe system at the Department of Earth Sciences, Cambridge. These

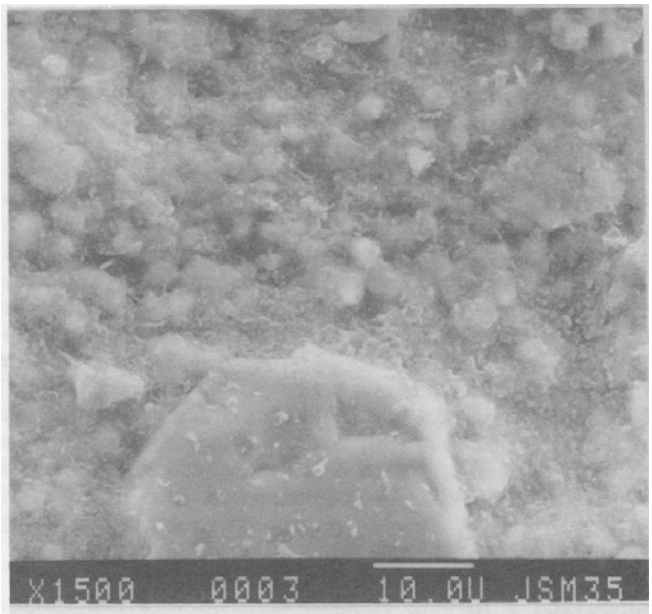

Figure 3. Scanning electron photomicrograph of apatite + garnet zone showing a euhedral garnet, hexagonal prisms of apatite and spiral wisps of graphite. Scale bar is 10 microns.

results indicate that the garnets are spessartine-almandine solid solutions with about 10% of the grossular and less than 2 % of the pyrope molecules present. Whilst the small garnets in the apatite and quartz + garnet zones appear to show a restricted range of chemical composition and little zoning the larger garnets at the margins of the lenses show strong compositional zoning in terms of the relative amounts of iron and manganese they contain.

Good analyses of the small garnets in the apatite and quartz + garnet zones of the lenses proved difficult to obtain because of the presence of inclusions of quartz, apatite and graphite. A typical apatite zone garnet analysis is given in Table 1 and shows approximately twice the amount of MnO to that of FeO to be present.

The central portions of the zoned garnets in the phyllosilicate layer close to the margins of the lens show approximately equal amounts of MnO and FeO although sometimes there is an excess of MnO but not to the same extent as that in the apatite zone garnets. However the margins of the zoned garnets show extreme preferential concentrations of MnO with up to three times the proportion of MnO to FeO. Means of several point analyses of centres and margins of two different garnets are shown in Table 1.

Two garnets were selected for more detailed analysis. Garnet A is 1.4 mm in diameter and occurs in the phyllosilicate layer within 2 mm of the margin of the lens. It shows a series of sectors of radially arranged inclusions. In crossed-polarised light repeated concentrically arranged slightly anisotropic zones may be picked out. The results of a series of point analyses across clear garnet sectors are plotted in Figure 4. Despite the repeated zoning observed optically the chemical zoning apparently

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		Zoned garnet A		Cross habit garnet B		
	Garnet from apatite zone	Centre(2)	Margin (2)	Centre (2)	Half-way (3)	Margin (4)
Weight % oxides:						
SiO_2	37.98	36.52	36.56	37.12	37.04	36.78
Al_2O_3	20.02	20.62	20.78	20.06	20.64	20.48
MgO	0.31	0.37	0.37	0.17	0.43	0.23
CaO	3.7	4.35	3.73	4.34	5.07	3.55
MnO	24.41	19.37	29.14	20.15	18.58	27.12
FeO	13.61	18.16	9.93	17.41	18.18	11.91
	100.23	99.39	100.51	99.25	99.94	100.07
Cation proportions (24 oxygens):						
Mg	0.075	0.091	0.09	0.042	0.105	0.056
Ca	0.642	0.771	0.651	0.767	0.883	0.626
Mn	3.347	2.712	4.019	2.813	2.559	3.78
Fe	1.842	2.51	1.352	2.4	2.472	1.639
	5.906	6.155	6.112	6.022	6.019	6.101
Si	6.147	5.989	5.953	6.065	6.023	5.97
Al	3.82	4.02	3.988	3.899	3.957	3.973
End member molecules (%):						
Pyrope	1.27	1.5	1.47	0.69	1.74	0.92
Grossular	10.87	12.67	10.65	12.73	14.68	10.26
Spessartine	56.67	44.57	65.76	46.72	42.52	61.95
Almandine	31.19	41.26	22.12	39.86	41.07	26.87

Table 1 Average chemical analyses of garnets from Treworld (Number of probe points averaged given in brackets)

shows a smooth reciprocal variation in Fe% and Mn% from the centre of the grain to the margin with the highest Mn% at the margins and the highest Fe% at the centre.

Garnet B is 0.9 mm in diameter and occurs at the margin of the lens. It has a very distinctive habit with cross shaped sectors of garnet separated by areas crowded with inclusions and both surrounded by an irregularly developed clearer rim to the grain (see Fig. 5). Point analyses along two arms of the grain and of the clear rim were made and these are plotted in Figure 5. It is apparent that the zoning here is more complex. Again overall the highest Mn% and lowest Fe% are found at the margin of the grain. However the central zone shows an initial decrease in Mn% before levelling out along the arms and then increasing at the margin. That is the inner portion of the grain shows the opposite trend to that of garnet A.

Compositional zoning in metamorphic garnets is common and may be a complex function of among other things, the rock composition and thermal history, the chemical reactions in which the garnet is involved and the oxygen fugacity (see Deer *et al.*, 1982 for a review). However, low to medium grade garnets often show simple bell-shaped distribution curves with centres relatively enriched in Mn and Ca and margins relatively enriched in Fe and Mg. The patterns have been widely interpreted as indicative of prograde metamorphic crystallisation whilst 'reverse' zoning with sharp marginal increases in Mn have been interpreted variously as the result of superimposed metamorphic events or as a retrograde metamorphic effect involving resorption of garnet.

The garnets described in this paper are more manganese rich than most considered in other studies of zoning and direct analogies of zoning behaviour are almost certainly inappropriate. The marked increase in Mn and reciprocal decrease in Fe towards the margins of garnets at the edge of the lens must be a reflection of changing conditions during growth. Solid-state diffusion at these temperatures is unlikely to have been important and the euhedral nature of the garnets together with their lack of replacement textures makes retrograde resorption of garnet equally unlikely. The physical conditions likely to give rise to the observed compositional profiles would include depletion in available Fe relative to Mn, and

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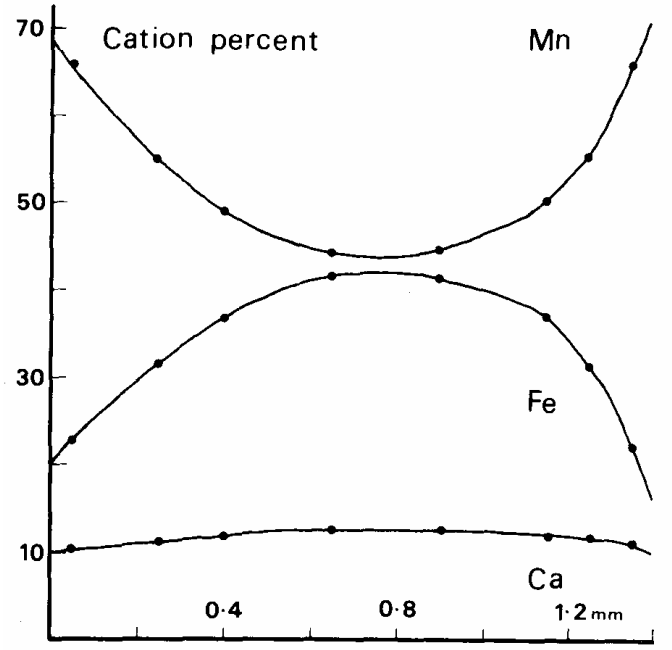


Figure 4. Plot of results of microprobe traverse across zoned garnet A. Cations expressed as a percentage of cation sum of (Ca + Mg + Fe + Mn).

decreasing temperature and increasing oxygen fugacity (Hsu, 1968) but it has not proved possible to select any one of these to be a more likely cause than the others. The complex profile of garnet B suggests a change in conditions of crystallisation and a superimposition of metamorphic events cannot be ruled out.

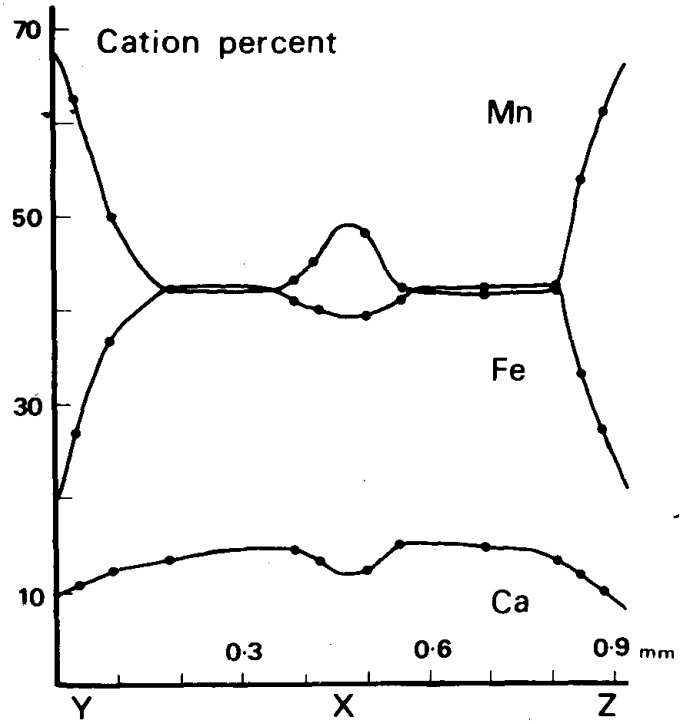


Figure 5. Plot of results of microprobe traverses across cross-habit garnet B. Cations expressed as a percentage of cation sum of (Ca + Mg + Fe + Mn).

Habit of the garnets

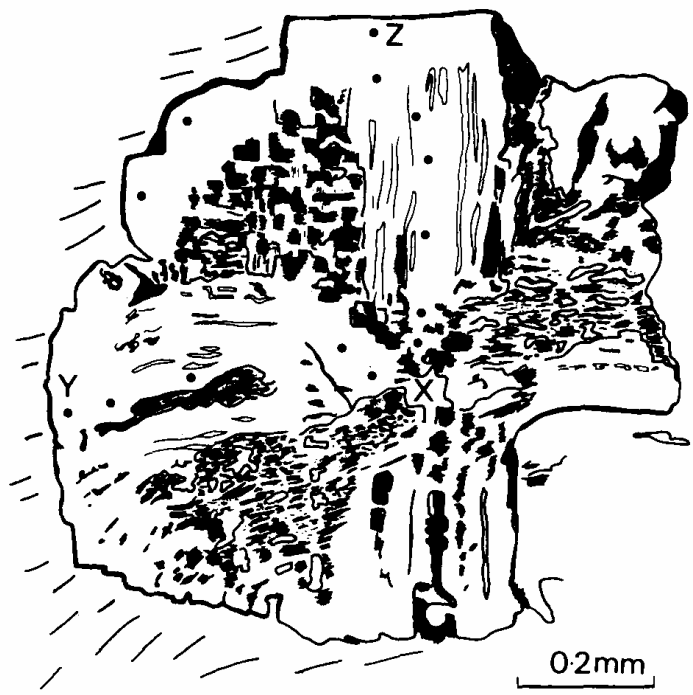
The distinctive habit of the garnets at the edge of the lens may be important in any explanation of the chemical zoning. Within the grains two types of sectorial growth may be distinguished (see Fig. 5). One type of sector shows incomplete garnet growth and many inclusions of quartz and graphite aligned parallel to the external fabric. The other type is characterised by much more complete garnet growth. These sectors have developed radially from the centre of the grain and are often terminated by well-developed crystal faces. They appear to represent directions of preferential crystal growth, They contain fewer inclusions and those that are present are usually quartz grains which are extremely elongated normal to the crystal faces and they may be orientated at high angles to the external foliation. The quartz must have grown at the same time as the garnet as a product of the garnet-forming reaction.

The rims of the garnets are usually clear and free of inclusions. This contrast suggests at least a change in growth rate if not a break in growth and may be the key to zoning patterns,

The manganoan garnets from Villapark are illustrated in plate 4 of Reid *et al.* (1910) and show very similar sectorial growth patterns.

Chemical composition of the shales of Treworld

Partial chemical analyses of five samples of the slates at



Sketch of garnet B with traverses X - Y and Y - Z and analysis points indicated.

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the Treworld exposure are presented in Table 2. Analysis was by automatic X-ray spectrometry and does not include determination of loss on ignition. Each sample is composed of fresh rock chips collected over a vertical height of two metres and thoroughly mixed. The analyses are therefore averages of the slate composition and not the precise composition of any particular layer. Sample 4603 was collected from closest to the garnetiferous nodules.

The samples all have low sulphur contents suggesting that they are unlikely to have been formed from anoxic, sulphur-rich black muds unless the sulphur has been subsequently removed. Sample 4603 shows very anomalous characteristics compared with the others. It is significantly lower in silica/and higher in iron, manganese and particularly phosphorus. The trace elements show a more erratic distribution and only copper is clearly enriched in sample 4603.

Discussion

Any explanation for the origin of the nodules must attempt to take into account all the following factors:-

(i) the presence of carbonaceous material;
(ii) the phosphorus concentration;
(iii) the concentration of Fe and Mn relative to the surrounding shales;
(iv) the apparent concentration of Mn relative to Fe in the nodules compared with the surrounding shales.

Table 2 Chemical analyses of chip samples of slate at Treworld

Sample No.	4643	4645	4603	3645	3667
SiO_2	52.16	66.1	43.65	57.14	60.68
Al_2O_3	22.89	15.89	15.87	20.93	22.97
TiO_2	1.04	0.91	0.59	0.61	0.53
Fe_2O_3	9.63	7.33	24.82	8.39	5.42
MgO	2.75	1.19	3.23	1.93	1.75
CaO	0.42	0.05	0	0.03	0.08
Na_2O	0.58	0.37	0.12	0.55	0.84
K_2O	4.74	3.53	2.79	4.25	4.84
MnO	0.22	0.17	0.76	0.31	0.09
P_2O_5	0.12	0.14	2.02	0.09	0.11
As	170	163	250	25	7
Ba	718	2076	1294	530	554
Co	15	16	47	15	7
Cr	156	128	88	124	136
Cu	36	63	237	30	22
Ni	66	66	37	61	26
Pb	29	30	28	21	33
Rb	206	127	96	184	219
Sr	155	83	51	70	88
S	-	246	231	197	489
V	153	266	147	159	232
Zn	117	125	162	101	61

Major elements in weight % oxide and trace elements in ppm element.
Sample 4603 was collected closest to the garnetiferous lenses.

These factors may have been features of the original sedimentary rocks or introduced either completely or in part during subsequent diagenetic and metamorphic processes.

It is difficult to envisage a purely metamorphic process to account for the present concentration of graphite and phosphorus and also rather unnecessary when sedimentary diagenetic processes are known that could account for these relationships. It is not unreasonable to assume that some of the observed variations in chemistry predate the metamorphism. Stanton and Williams (1978), for example, concluded that at Broken Hill, New South Wales, a very delicate chemical pattern of original sedimentation had been preserved despite high grade metamorphism.

A key factor in accounting for the nodules seems to be the central zone of apatite. Manheim and Gulbrandsen (1979) in a review on marine phosphorites state that many phosphorites are formed diagenetically under reducing conditions from phosphate rich pore waters in anoxic sediments rich in organic debris. Thus if our original sediment contained organic debris this could have been reduced to carbonaceous material and also contributed phosphorus to the pore waters. The phosphorus may have first formed carbonate fluorapatite which upon metamorphism would have been converted to apatite (Matthews and Nathan, 1977). Alternatively some or all of the phosphorus may have initially combined with iron to form an iron phosphate.

The behaviour of iron and manganese is more difficult to define. They could have initially been concentrated in the sediments within volcanic debris or as the result of hydrothermal hot spring activity on the sea floor. The possibility of primary manganese haloes being preserved around metamorphosed stratabound base metal deposits has been discussed extensively by Stumpfl (1979) and if hot spring activity was the prime cause of the manganese concentrations then the distribution of manganese garnets throughout the schists would be of special interest to mineral prospectors. It appears, however, that the iron and manganese distribution is related to the nodules and if the phosphorite formation is likely to have been diagenetic then the present concentration of iron and manganese around the apatite must be of diagenetic or later origin.

The diagenetic movement of both manganese and iron through sedimentary piles has been well documented (for a review see Roy, 1981) and provides a potential mechanism for the preferential concentration of one or other or both elements. In essence the underlying principles are simple, in practice the exact mechanisms may be very complex. The reduced forms of the iron and manganese cations, Fe^{2+} and Mn^{2+} are more readily taken into solution than the oxidised forms, Fe3+ and Mn4+, In a reducing environment, such as a sediment rich in organic material, the oxidised forms will tend to be reduced and removed from the sediment into the pore waters. With consolidation the pore waters will migrate upwards and if an oxygenated layer of sediment is present close to the sediment-water interface, iron and manganese oxides or hydroxides will be precipitated. If no oxygenated

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sedimentary layer is present the pore waters will contribute iron and manganese into the overlying bottom waters. Thus there is a potential 'zone refining' mechanism available to progressively enrich the oxygenated layer of sediment with iron and manganese with thickening of the sedimentary pile. This mechanism could also work against concentrations of these elements being preserved in the sedimentary pile. This is one of the reasons why ocean floor manganese nodules are not commonly preserved in the fossil record. Elaborations of this basic mechanism will depend on the exact chemistry of the system. Iron may be preferentially retained in the reduced sediment as pyrite following reaction with hydrogen sulphide and the presence of bicarbonate ions in the pore waters may lead under suitable conditions to the precipitation of carbonates. Bacterial action may also have an important influence on the distribution of iron and manganese. The actual form of iron and manganese minerals in sediments and on the sea floor is often as encrustations on material which has acted as a substrate upon which they are able to grow.

It seems probable, therefore, that the phosphatic nodules acted as nuclei for the diagenetic deposition of iron and manganese minerals. Unfortunately the subsequent metamorphism has removed any possible evidence of the exact nature of these minerals. However, Roy (1981, p. 115) points out that in manganese mineral deposits Mn-oxide and silicate-oxide assemblages very rarely occur in association with carbonaceous metasediments whereas Mn-carbonates formed from interstitial pore fluids under reducing conditions are found in association with carbonaceous metasediments. Possibly the diagenetic precursors of the metamorphic spessartine-almandine garnets were iron and manganese carbonate minerals growing around the phosphorite nodules.

If the nodules are of diagenetic origin then metamorphism of these nodules would account for the apatite + garnet cores and the garnet rich rims. However the mode of formation of the quartz + garnet zones is problematical and whether the quartz rich zones reflect a diagenetic feature or result from metamorphic differentiation is not clear.

Within the nodules the availability of aluminium may have been limited and the restricted size of garnets there probably indicates only limited diffusion of aluminium into the nodules. Because of the limited supply of materials the time period of garnet growth within the nodules may have been shorter than that for the garnets at the rim where Fe, Mn and A1 were more readily available and this may be a factor in the development of zoning in the larger garnets at the rim of the nodules. The minerals associated with the garnets may also be an important factor in the development of the garnet zoning. It is only at the margin of the nodules that garnet is associated with chlorite and further work is needed to study the Fe and Mn partitioning between garnet and chlorite.

The recognition of probable former phosphatic nodules coated with iron and manganese concretions through the examination of these garnetiferous nodules suggests that a much more extensive study of the geochemistry of the rocks of Upper Devonian and Lower Carboniferous age in the area could yield more information about their environment of deposition and their diagenesis despite subsequent metamorphism. A limited number of chemical analyses of the Transition Group Slates have been given here and a possible correlation between anomalous phosphorus in the shales and the identification of phosphatic nodules has been suggested. Other unpublished analyses we have seen of the slates, indicate other anomalously high phosphorus contents and suggest that phosphatic nodules may occur elsewhere in the Transition Group.

Extensive diagenetic changes in the chemistry of the shales could be instrumental in the formation of economic mineralisation.

Conclusions

Spessartine-almandine garnet bearing nodules discovered in graphitic shales near Boscastle, north Cornwall contain apatite rich cores. It is proposed that phosphorite nodules were formed diagenetically from pore waters in organic-rich anoxic muds and then acted as nuclei for diagenetic concretions of iron and manganese minerals, possibly carbonates rather than oxides. The zoning of these diagenetic features has been preserved despite subsequent metamorphism by the formation of apatite + garnet cores and garnet rich rims to the nodules. Further study of the chemical variation in the sedimentary rocks in the north Cornwall area could yield useful palaeo-environmental information.

The garnets at the edge of the nodules offer a new example of 'reverse zoning' with their relative enrichment in Mn and depletion in Fe towards the margin of the crystals. The growth habit of these garnets suggests that a change of growth rate may have been an important factor in the development of this zoning. However, the influence of a superimposed metamorphic event cannot be ruled out.

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Read at the Annual Conference of the Ussher Society, January 1984

The Pleistocene Chadbrick River gravels of the Cary Valley, Somerset: Amino-acid racemisation and molluscan studies.

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D. D. GILBERTSON
N. M. THEW

Hunt, C. O., Gilbertson, D. D. and Thew, N. M. 1984. The Pleistocene Chadbrick river gravels of the Cary valley, Somerset: amino-acid racemisation and molluscan studies. *Proceedings of the Ussher Society, 6*, 129-133.

Cemented Pleistocene river gravels, here termed the Chadbrick Gravels, of a former River Gary have been exposed slightly downslope of a marked terrace landform at Hurcott Farm, Somerton, Somerset (ST5125 2955). The Chadbrick Gravels form the most important part of a sequence of five gravel units in the Valley. Field mapping, coring and the excavation of trial pits in the Chadbrick Gravels has revealed epsilon cross-bedding suggesting deposition in a meandering stream or river. Subsequent studies of the molluscan fauna suggested the presence of a major river surrounded by woodland in an interglacial environment. Amino-acid racemisation studies of a valve of Corbicula fluminalis yielded a ratio of 0.18 which is broadly similar to those of freshwater molluscs from the Burtle Beds of the Somerset Levels. These ratios suggest that these molluscs should be attributed to amino-acid stage II, "broadly" the Ipswichian interglacial. Petrological studies of the Chadbrick Gravels suggest that at this time the headwaters of the River Yeo drained through the Gary Valley and had not yet been captured by a tributary of the River Parrett. Unresolved problems still remain concerning the age and origins of other molluscan fa unas from the Burtle Beds as well as the age(s) and environmental significance of the other gravel units in the Gary Valley. No evidence has been recorded which would suggest the area has been glaciated.

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Introduction

This paper records the sedimentary structures, molluscan palaeocology, and amino-acid racemisation dating of the fossiliferous Pleistocene river gravels, termed the Chadbrick Gravels, located near the confluence of the Cary and Chadbrick valleys near Somerton, Somerset (Fig. 1). These Chadbrick Gravels provide a reference from which to build a Pleistocene stratigraphy of this area which is currently poorly known (Hunt, 1985). This is unfortunate because, although glaciation is not the subject of this report, the Cary valley lies in a critical location for resolving the problems of the Quaternary Geology of the south-west of England (see Kidson and Tooley, 1977; Stephens, 1970; Gilberston and Hawkins, 1978).

In 1954 the Reverend J. Fowler of Somerton presented a fragment of cemented gravel containing freshwater molluscs to the British Museum (Natural History). The molluscan fauna of the gravel was recorded by Gilbertson and Beck (1975), who noted the presence of the freshwater, thermophilous bivalve *Corbicula fluminalis*, the interglacial affinities of the fauna, and estimated that it was obtained from a well developed terrace landform in the vicinity of Hurcott Farm, Somerton, some 20-30 m above the modem floodplain of the River Cary. Neither the exact source of the gravel fragment or a reliable estimate of its age could be made.

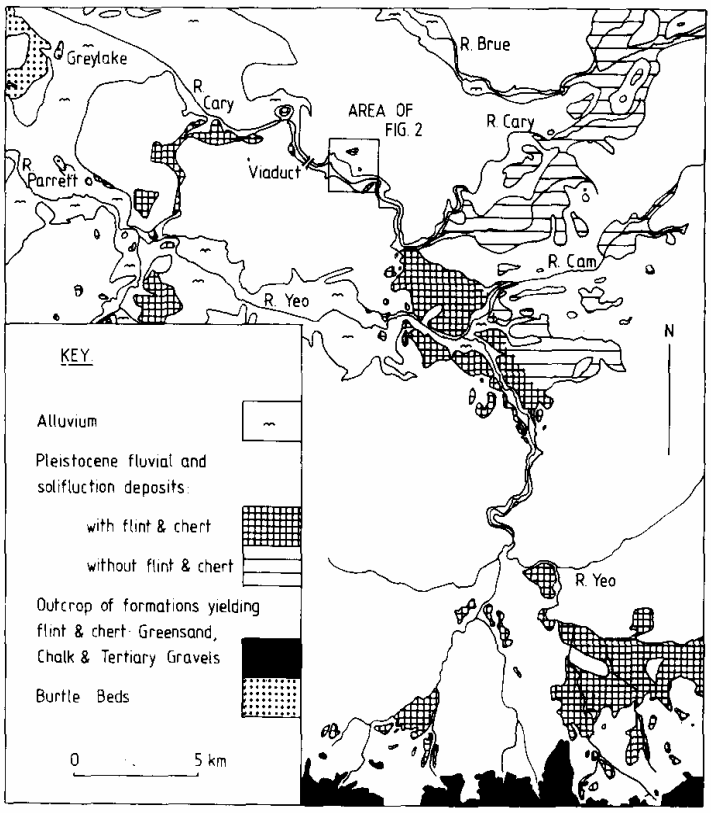

Figure 1. Geological map of study area.

In the course of field mapping 1979-1982, similar, but

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more weathered, fragments of' cemented gravels were found where they had been brought to the surface by deep ploughing at ST51252955, slightly below the front of the terrace landform at Hurcott Farm (Fig. 2). A borehole survey using a hand auger was combined with the excavation of a line of six trial pits downslope across the terrace and the scatter of gravel fragments in order to assess the dimensions and field relationships of the gravel body (Fig. 2).

Stratigraphic relationships

These field studies revealed the distribution and stratigraphic relationships shown in Figure 1. Five stratigraphically distinct deposits were noted:

1) The highest deposits occur at 43 m O.D. overlying the terrace landform at Hurcott Farm. They comprise 0.4 m of very coarse, poorly sorted gravel with a cold-stage molluscan fauna with *Pupilla muscorum L.* and *Trichia* cf. *hispida*, which is suggestive of a Pleistocene cold stage.

2) Six metres downslope from the Hurcott Farm terrace landform at 37 m O.D. heavily cemented gravels were found. They are relatively well preserved and cemented to the underlying Rhaetic limestone. Their sedimentology, palaeontology and correlation are described in the next sections. They are termed the Chadbrick Gravels after the small valley in which they were exposed.

3) at 29 m O.D., some 8 m below the lowest outcrop of the Chadbrick Gravels, terrace features were located upon a unit of plane-bedded, fine reddish sands and gravels. No macro- or microfossils were recovered from these deposits.

4) Two metres above the modern Cary floodplain, a terrace surface was located at 14 m O.D., developed on coarse, unfossiliferous gravels. The unit is currently known only from borehole studies.

5) Woodward (1905) noted a further gravel unit underlying the modern Cary floodplain in foundation trenches for the railway viaduct (ST492292).

The field relationships of these deposits suggest that they form a progressively older sequence, the valley bottom gravels noted by Woodward (1905) beneath the modem floodplain alluvium being the youngest of the gravel units, the highest and oldest resting on the Hurcott Farm terrace. At present there are insufficient other data to verify this relative chronology. The mapping study has shown that the fossiliferous, cemented gravel fragment recorded by Gilbertson and Beck (1975) was probably not obtained from the Hurcott Farm terrace, but rather from a stratigraphically younger, and altitudinally slightly lower location. The relationships of these units to those recorded previously from the valley by Woodward (1905), and Geyl (1974) are discussed in Hunt (1984). The field mapping reported here did not record any deposits suggestive of past glaciation of the region (Hunt, 1984).

Sedimentology

The Chadbrick Gravels comprise a 0.25 - 0.35 m thick sequence of clast-supported, imbricated, epsilon cross-bedded, cemented gravel. One incomplete set of epsilon

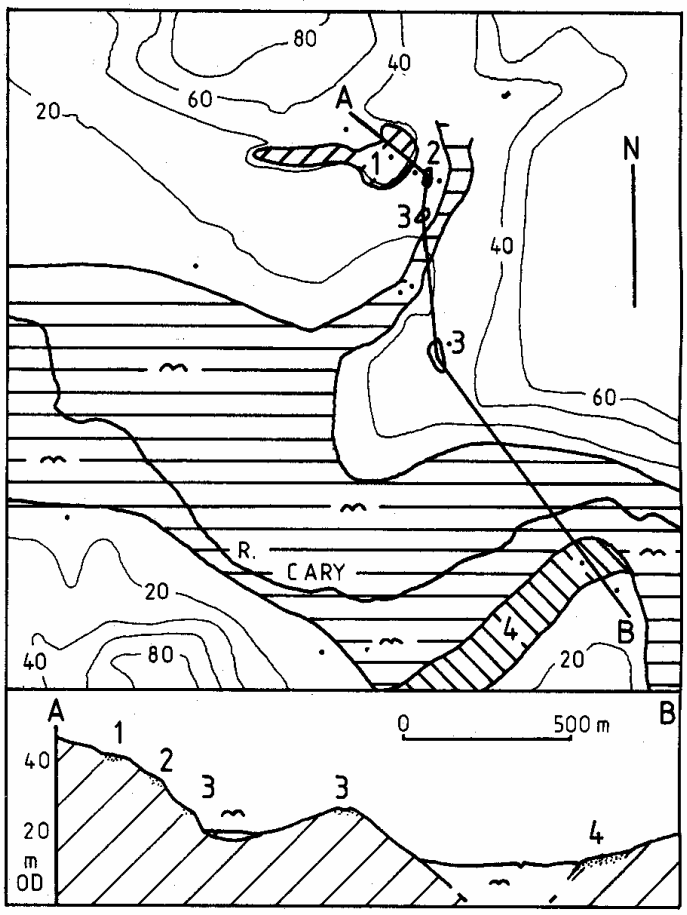

Figure 2. Map and section of detailed study area.

cross-beds were present, the top of the unit having been removed by erosion, it is estimated from the remnant sedimentary structures that the original height of the cross-beds was in the region of 0.4 m, suggesting a channel depth for the former river of at least that depth. Epsilon cross-bedding is commonly associated with meandering streams (Reineck and Singh, 1975). Visual examination of these cemented gravels indicates they are well sorted, with an estimated clast size of 5.0 mm. The clasts are uniformly well-rounded. Examination of the surface of the cemented gravels yielded a Krumbein (1941) index average of 6.6. The high degree of textural maturity shown by these gravels suggests deposition in a rather stable fluvial environment. The sedimentary structures present suggest a point bar in a meandering river channel.

The molluscan fauna

The firmly cemented nature of the Chadbrick Gravels also frustrated their disaggregation for molluscan studies. Therefore only molluscs on the faces of the blocks could be identified. Despite this problem numerous small and potentially fragile taxa were noted. However, the inability to inspect most of the voids between the clasts into which many small shells might have entered as an infiltration deposit suggests the relative proportions of these smaller taxa are under-represented in Table 1.

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Oxyloma cf. pfeifferi	5
Cochlicopa cf. lubrica	1
Vallonia cf. pulchella	1
Discus rotundatus (MULLER)	1
Clausiliidae	1
?Helicella sp.	1
Trichia cf. hispida	6
Valvata cristata (MULLER)	1
Valvata piscinalis (MULLER)	88
Bithynia tentaculata (L.)	24
Physafontinali, (L.)	1
Lymnaea stagnalis (L.)	1
Lymnaea peregra (MULLER)	41
Planorbis spp.	2
Gyraulus laevis (ALDER)	7
Ancylus flutiatalis (MULLER)	1
Unio sp.	3
Corbicula fluminalis (MULLER)	18
? Sphaerium sp.	1
Pisidium amnicum (MULLER)	9
Pisidium clessini Neumayr	5
Pisidium henslowanum SHEPPARD	3
Pisidium nitidum JENYNS	1
Pisidium subtruncatum MALM	4
Pisidium spp.	22

Table 1. Molluscan fauna of the Chadbrick Gravels, near Hurcott Farm, Somerton, Somerset. Nomenclature follows Ellis, 1962; Kerney, 1976; Walden, 1976. Figures indicate number of specimens identified.

Two hundred and forty eight specimens were identified and attributed to twenty five taxa from examination of twelve fragments of cemented gravel weighing approximately 5 kg. (Table 1).

Palaeoenvironments

The molluscan fauna is dominated by aquatic taxa, especially an assemblage which characteristically might occur in riverine deposits (67%). This comprises the large freshwater bivalves *Corbicula fluminalis* and the undifferentiated *Unio* spp; the small bivalves *Pisidium amnicum*, *P. henslowanum*, *P. nitidum*, *P. subtruncatum*, the extinct *P. clessini;* and *Valvata piscinalis, Bithynia tentaculata,* and *Ancylus fluviatilis.*

Comparisons with previous studies of Pleistocene riverine assemblages by Kerney (1971), Gilbertson (1980) and Briggs and Gilbertson (1980), together with modern ecological surveys (e.g. Okeland, 1964; 1969; 1979) suggest that two major facies groups are represented. *Bythinia tentaculata, Valvata cristata, Lymnaea peregra and Physa fontinalis* all prefer slow-moving, calcareous, eutrophic, well-vegetated reaches; the prosobranchs *Bythinia tentaculata, Valvata piscinalis,* and *Valvata cristata* requiring soft; fine-grained substrates. *Valvata piscinalis* also prefers quiet, fairly deep water, *Pisidium subtruncatum, Pisidium nitidum, Unio* spp., and *Sphaerium* spp., all require well-oxygenated, eutrophic water, with locational preferences close to this facies group.

The second group comprises *Ancylusfluwatilis, Pisidium amnicum* and *Pisidium henslowanum* which are normally characteristic of faster flowing reaches with coarser sandy substrates and relatively little aquatic vegetation. *Corbicula fluminalis* (extinct in Europe) and *Pisidium clessini* (totally extinct) may also be in this group.

Two "generalist" species, *Lymnaea peregra* and *Gyraulus laevis*, can be found in both types of environment.

The occurrence of these two facies groups, with the former predominant, is diagnostic of a meandering river, with slow-moving, deeper water in "pools", and faster flow over the "fifties" of the fiver or stream bed. Hard water conditions are indicated by the several calciphiles present.

The terrestrial species include one characteristic marsh taxon, *Oxyloma* cf. *pfeifferi*, and several taxa characteristic of damp, sheltered habitats. *Discus rotundatus* is normally associated with wooded or scrubland habitats, suggesting a former tree cover adjacent to the river.

Many of the taxa noted are climatically very tolerant; however, Corbicula fluminalis and Discus rotundatus are only known from interglacial or warmer interstadial contexts (see Kerney, 1977).

The composition of this fauna provides no reason to believe that any components of it have undergone anything other than the normal pene-contemporaneous re-working, transportation and re-deposition that is associated with the accumulation of molluscan death assemblages in rivers. These processes are becoming increasingly well known and to some extent quantified by field and laboratory studies (see Briggs, Gilbertson and Harris 1984 a and b). consequently it is concluded that the molluscan evidence indicates that the Chadbrick Gravels collected in a temperate or at least mild climate as a result of deposition in a meandering river or stream bordered by marsh, and possibly wet grassland with trees or scrub lining or over-hanging the river.

Dating and Correlation

The molluscan evidence provides only limited evidence concerning the correlation and dating of the Chadbrick Gravels. The fauna is essentially a Pleistocene warm stage riverine facies fauna. Both *Corbicula* and *Pisidium clessini* have been extinct in the British Isles since the early Devensian. *Corbicula* tends to be rare (Kerney, 1971) in deposits attributed to the Hoxnian warm stage as defined in Mitchell, Penny, Shotton and West (1973). *Nematurella runtoniana* which is known in essentially similar deposits of Cromerian age in Britain (Kerney, 1977; Gilbertson, 1980) is not present, although this could also be a result of small sample size, chance, or being missed in the calcite cement of the gravels. Unfortunately there is very limited information on Pleistocene molluscan biostratigraphy in the south west of Britain.

Several Pleistocene freshwater molluscan assemblages of similar composition are known from Pleistocene deposits closer to the present Severn Estuary. These include the Burtle Beds downstream of the River Cary (Fig. 1) which have yielded non-marine Molluscs as well as inter-tidal assemblages (Jackson and Bulleid, 1931; Bulleid and Jackson, 1938, 1941; Kennard, 1941; Kidson *et al.*, 1978, 1981; and Gilbertson, 1979) and the complex of freshwater and inter-tidal deposits at Kenn, north of

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Mendip (Figure 1; Gilbertson and Hawkins 1978). Unfortunately these are all facies faunas and offer no reliable biostratigraphic correlation between these deposits.

Aminostratigraphy

Amino-acid racemisation ratios for one valve of Corbicula fluminalis from the Chadbrick Gravels at Hurcott Farm, and two specimens from the Burtle Beds (supplied from the collections of the British Museum (Natural History) by Dr. M. Cooper) were determined at INSTAAR, University of Colorado, Boulder, by K. H. and D. C. Davies using standard techniques (as used in Davies, 1983). The ratios are given in Table 2.

LOCALITY	BM(NH) code.	INSTAAR lab. code no.	al:il ratio
Chadbrick Gravels; Hurcott Farm. (ST 5125 2955)	LL 2067	AAL 2382A	0.18
Burtle Beds at:			
Old Sea Bank (ST3933 see below)	LL2503	AAL 2382B	0.18
Othery (ST3933 see below)	LL18231	AAL 2382C	0.26

Table 2. Amino-acid racemisation ratios from *Corbicula fluminalis* from the Chadbrick Gravels, Hurcott Farm, Somerton, Somerset; and the Burtle Sand Beds of Somerset at "Old Sea Bank" and Othery, (Figure 1). "Old Sea Bank" is probably "Othery Old Sea Bank" in the Carnelly Collection, obtained from Greylake (ST 3933) (Gilbertson, 1979). "Othery" is probably "Othery/Sedgemoor Old and New Pits"—probably again the Greylake site (Gilbertson, 1979).

The Chadbrick Gravels and one of the Burtle sites yield similar ratios of 0.18 which places them in the Amino-acid Group 2 of Miller, Hollin and Andrews (1978), which they regard as representing the Ipswichian interglacial. The Othery site yielded what appears to be a notably different ratio. Its significance is unclear. It may represent a genuinely older deposit, a different thermal history in the Othery sediments to that experienced at the "Old Sea Bank", or a reworked older shell as suspected in several raised beach deposits investigated by Davies (1983) from south-west Britain. Further work is needed to resolve these problems. Nevertheless, these data suggest the Chadbrick Gravels should be referred to the Ipswichian interglacial and correlate with at least part of the freshwater component of the Burtle Beds.

Palaeogeography

Petrographic analyses of the Chadbrick Gravels are presented in Table 3. These data provide evidence on the drainage system of the time. The Chadbrick Gravels are seen to be dominated by clasts derived from the local Rhaetic and Liassic bedrocks. Significantly, pebbles of flint, Greensand chert and sandstone are also present. In samples obtained from the Upper Cary Valley these rock types are absent; published and unpublished mapping (see Hunt 1985) has indicated that there are no Cretaceous or Tertiary outcrops in the Cary catchment which could have yielded them. Neither has unpublished mapping recorded in Hunt (1985) located Pleistocene drifts which might be their source. However, Cretaceous rocks do outcrop south of Yeovil in the Yeo catchment and flint and Greensand chert and sandstone are found in the Yeo valley gravels (Table 4).

These data suggest:
1) that the proto-Cary included the present headwaters of the River Yeo within its catchment during the period represented by the Chadbrick Gravels;
2) that at an unknown time after the deposition of the Chadbrick Gravels near Hurcott Farm, the headwaters of the River Yeo were captured and diverted into their present alignment as a tributary of the River Parrett, hence cutting off the River Cary from sources of flint, Greensand chert and sandstone. Presumably this occurred during the Devensian cold stage in the low ground south of the Langport-Somerton ridge.

Conclusions

The Chadbrick Gravels exposed near Hurcott Farm, Somerton, were deposited by a meandering river with a rich aquatic fauna of thermophilous molluscan taxa. The river was bordered by marsh and wet grassland, possibly with trees or scrub nearby, The molluscan assemblage indicates deposition in a warm interglacial climate.

LITHOLOGY	PERCENTAGE
Calcareous siltstone	33.6
Micritic limestone	29
Oolitic limestone	15.9
Flint/Chert	10.3
Bioclastic limestone	3.7
Ironstone	2.8
Red mudrocks	1.9
Liostrea and other fossils	1.9
Greensand	0.9

Table 3. Petrographic analysis of the Chadbrick Gravels (5-20 mm fraction)

CATCHMENT	UNIT	%FLINT/ CHERT	%GREEN-SAND
Upper Yeo		48.35	present
Middle Yeo		12.23	2.07
Upper Cary	-	-	
Lower Cary	1	0.7	-
	2	10.3	0.9
	3	22.4	present
	4	5.9	0.4

Table 4. Average percentages of Cretaceous and Tertiary lithotypes in the drift deposits of the Yeo and Cary catchments.

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Amino-acid racemisation studies indicates they may be correlated with the Amino-acid stage 2 of Miller et al. (1978), broadly equated with the Ipswichian interglacial. The Chadbrick Gravels may be locally correlated by aminostratigraphy with some, but not all freshwater molluscan assemblages from the Burtle Beds. The Chadbrick Gravels were deposited in a proto-Cary which may have drained the headwaters of the present River Yeo, as well as its own catchment.

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Read at the Annual Conference of the Ussher Society, January 1984

Structural models of the geology of the north Cornwall coast; a discussion

E. B. SELWOOD
J. M. THOMAS

Selwood, E. B. and Thomas J. M. 1984. Structural models of the geology of the north Cornwall coast; a discussion. *Proceedings of the Ussher Society, 6*, 134-136.

The existence of a southward directed overfold, the subject of much structural modelling, at the southern margin of the Culm Synclinorium of central south-west England is questioned. New evidence points to the existence of a large scale nappe structure underthrusting the main Upper Carboniferous flysch basin of the region.

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The geology of the north coast of Cornwall is attracting increasing interest as a basis for tectonic modelling, and conclusions are being drawn which are held to be of wide application elsewhere. These papers accept as a fact the existence of a transition from upright close folds in the north, to tight recumbent folds in the south with a southward tectonic transport direction (Fig. lc); a structure referred to as the "South Culm Overfold" (Ferguson and Lloyd, 1982) and the "Millook Nappe' (Rattey and Sanderson, 1982). It is generally agreed that the coast section is much dissected by north dipping faults, and most (e.g. Hobson and Sanderson, 1983) accept that the Rusey Fault Zone (Fig. lb) separates upright to overturned south-facing folds to the north from recumbently folded strata of higher metamorphic grade to the south which are continuous with the Tintagel High Strain Zone. However it is not always appreciated that a major overfold cannot be directly observed, and that it is itself a model (Freshney, 1965) based on the simplest interpretation of the facts available at the time of its formulation.

The overfold model gained considerable support by comparison with a similar model which had been proposed (Dearman, 1968) for the structures around the western margin of the Dartmoor granite some 50 kms to the east along the strike. The disposition of the strata west of Dartmoor previously interpreted as the recumbent limb of the overfold has now been shown (Isaac et al., 1982) to be due to large scale thrusting with a northwesterly transport direction. Similar structures involving northward transport of thrust sheets have been demonstrated westwards from Dartmoor to within 15 kms of the coast sections (Isaac et al., 1983; Stewart, 1982, Turner, 1982). It is now argued (Selwood, Thomas and Stewart, in press) that these structures persist even farther westwards to the north Cornwall coast; it follows that the overfold model as currently envisaged is no longer valid.

Hitherto it has usually been assumed that the rocks constituting the overfold from Boscastle north to Millook belonged to the Upper Carboniferous Crackington Formation; a thick sequence of mud cemented, frequently well graded, distal sandstone turbidites. Within the outcrop zone of the sandstones south of the Rusey Fault, thin fault-bounded slices of greyish-green slates of Lower Carboniferous age have been recognised as the Buckator Formation (Freshney *et al.*, 1972). It is now possible to demonstrate (Selwood, Thomas and Stewart, in press) that this formation also fields shallow water fades conodonts which extend its age into the Lower *velifer* zone of the Upper Devonian, and that it interfingers with the enclosing arenites which inland have yielded Lower Carboniferous faunas. Further, a reappraisal of the sedimentology of the "Crackington Formation" between Boscastle and the Rusey Fault indicates a shallow marine depositional environment, an interpretation that is consistent with its lateral equivalence to the Buckator Formation, but in direct contrast to that of the basinal turbidites of the Crackington Formation north of the Rusey Fault. These shallow water arenites and slates warrant separate formational status, and we propose to revive Boscastle Formation (Ashwin, 1958).

The Rusey Fault zone now assumes increased importance, for not only does it separate rocks of different structural style and metamorphic grade, but also successions of contrasting facies and ages.. The effect of this is to weaken considerably the concept of a simple overfold model.

It is now possible to demonstrate that the eastward projection of the Rusey Fault zone (Figure lb and beyond) limits the outcrop of the Crackington Formation s.s. to the north against a shallow water Lower Carboniferous succession to the south for a distance of over 13 kilometres. At the eastern limit of their outcrop these Lower Carboniferous rocks form part of a large scale nappe over-riding Tredorn Slate which crops out continuously to the coast south of the Lower Carboniferous sandstones (Fig. lb). Although the thrust is cut out southwards by a late high angle fault which extends through to the coast at Wlllapark,

Figure 1 North Cornwall Coast Geology
a. Location map of the area involved in South west England.
b. Locality map of North Cornwall Coast, with area of Upper Devonian--Lower Carboniferous sandstones marked as "Boscastle Formation".
c. Previous interpretation of the structural relationships of the coast section with late faults and normal faults restored (From Sanderson 1971, Fig. 2
d. Present proposals for the revised structural interpretation.
R-Rusey Thrust zone. B-Bodmin Moor. D-Dartmoor

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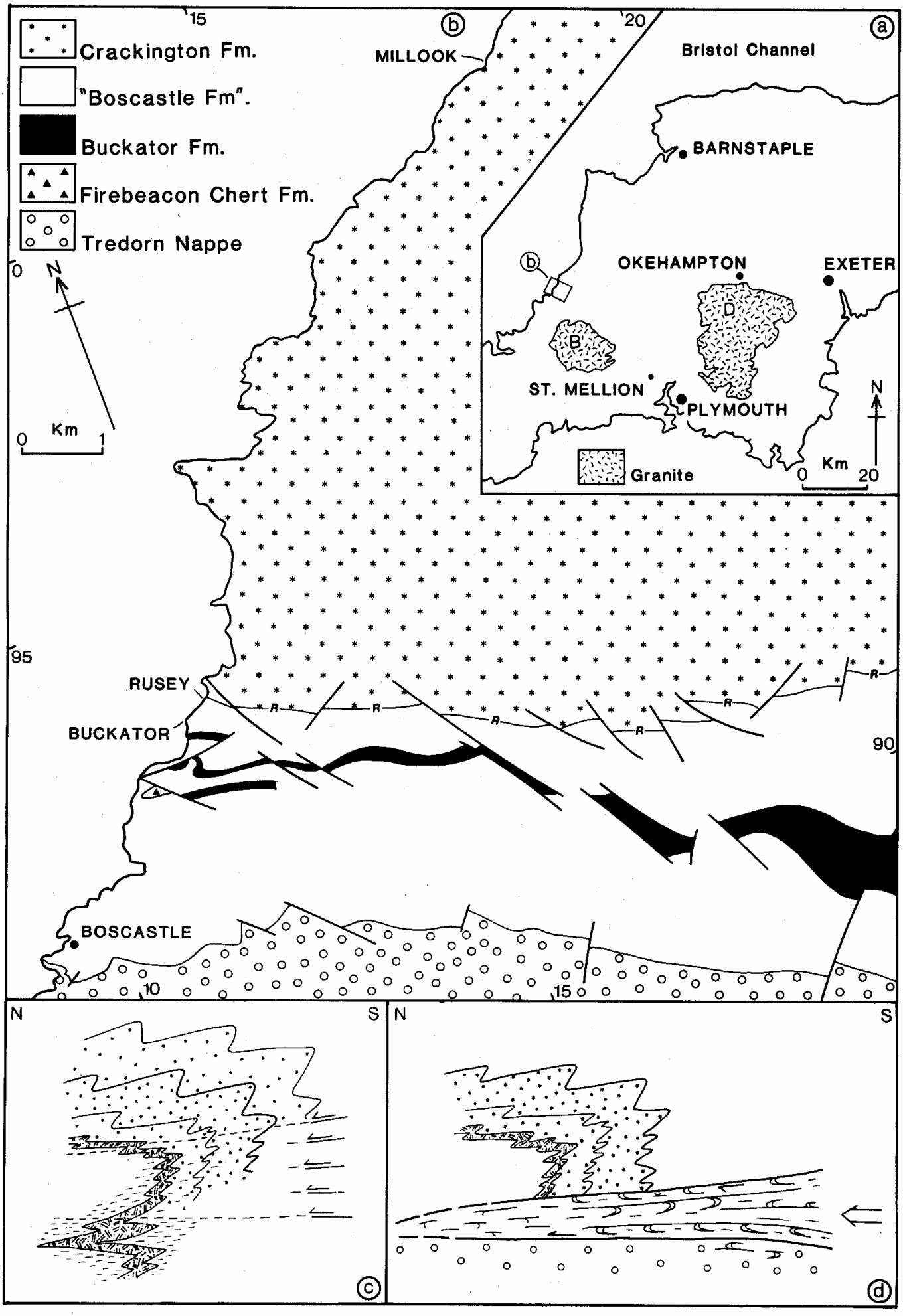

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there is no reason to doubt the continuity of the nappe. Thus the recumbent limb of the "South Culm Overfold" constitutes an integral part of the nappe structure; a view supported by the previously unrecorded occurrence of considerable thicknesses of blastomylonitic and phyllonitic fault rocks in gently north-dipping zones in the coast sections.

This reinterpretation of the structure necessitates the reversal of the southerly movement demanded by the existing overfold model. Inland all the evidence supports this view and even on the coast sections where southward transport was originally proposed, evidence for south facing is meagre or ambiguous (Rattey and Sanderson, 1982). North of Boscastle, not only are the sandstones of the north-south aligned sheath folds so strongly deformed that younging is difficult to prove, but also the folds themselves are frequently dismembered by shearing. East and west closing elements of sheath folds are common, but north or south closures are rare and their facing cannot be of great significance in such a complex structural setting. The regional geology however gives massive support for northward transport and allows the restoration of northward movement (Wilson, 1951) of the thrust pile south of Boscastle.

This presence of a large scale south-facing, fault dissected, overfold in the Crackington Formation (*sensu stricto*) north of the Rusey Fault as described by earlier authors, is fully supported. The early structure of the Rusey Fault appears to have been a north-dipping thrust (Freshney et al., 1972) which we interpret as an underthrust where the northern tip of the nappe descended beneath the southern edge of the main flysch basin of south-west England (Fig. ld). We interpret the south-facing overfold north of Rusey as a local southward transporting backfold developed in response to this northward underthrusting. The tectonic relationship at the Rusey Fault is thus comparable to that suggested by Zwart (1964) of a northward underthrusting infrastructure giving southward movements in the superficial rocks.

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Shallow-water Dinantian sediments in south-east Cornwall

M. J. WHITELEY*

Whiteley, M.J. 1984. Shallow-water Dinantian sediments in south-east Cornwall. *Proceeds of the Ussher Society, 6*, 137-141.

Three allochthonous Dinantian formations are recognised in south-east Cornwall overthrusting a monotonous sequence of Upper Devonian slates. The Brendon and Cotehele Formations are developed within the Greystone Nappe and the Crocadon Formation in the overlying Blackdown Nappe. The Crocadon Formation, a predominantly clastic, shallow-water, deltaic facies, is described in detail. This new interpretation establishes shallow-water sedimentation as an important aspect of Dinantian geology in central south-west England.

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Introduction

In a previous contribution that summarised the geology around St. Mellion, south-east Cornwall, the presence of a series of allochthonous thrust slices was established (Whiteley, 1982). These structural units form part of a thrust and nappe terrain that is widely recognised in south-west England (e.g. Isaac *et al.,* 1982; Shackleton *et al.,* 1982), the elements of which are derived from the south. The lowest structural unit, possibly representing autochthon, consists of a thick sequence of Upper Devonian slates (Kate Brook Slate Formation) and it is overlain by the Greystone and Blackdown Nappes (Isaac *et al.,* 1983). In the area between Bodmin Moor and Dartmoor the Greystone Nappe is widespread and it includes the Brendon and Cotehele Formations; these formations are dominated by shales and sandstones respectively but a lateral gradation between them has been demonstrated at Cleave (Isaac *et al.,* 1983, Fig. 12).

The Blackdown Nappe is thrust directly over the Greystone Nappe in south-east Cornwall. Here the sole constituent of the Blackdown Nappe is a thick sequence of clastic sediments termed the Crocadon Sandstone formation and although some of its characteristics have already been described (Whiteley, 1981) the interpretation of the Crocadon Formation in terms of shallow-water sedimentation is anew concept.

Crocadon Sandstone Formation

Definition

A sandstone dominated succession comprising feldspathic and micaceous sandstones with interbedded siltstones and shales. The sandstones are characteristically poorly sorted, weakly graded and < 0.75 m thick whilst the siltstones are well laminated and may display cross-bedding. Fossils other than plant material are infrequent but include goniatites, ostracods, conodonts and trace fossils. Exotic blocks of Famennian shale and Viséan chert are incorporated within various parts of the formation.

Type locality

A large disused quarry at SX39206575 in Crocadon Wood, 250 m due south of Crocadon Farm, St. Mellion, south-east Cornwall. A simplified log of the inverted sequence here is given in Isaac *et al.,* (1983, Fig. 13c).

Distribution and thickness

This formation is the major allochthonous unit in the region, occupying about 22 sq. km of high ground south of Callington. To the north and west it is thrust over varied lithologies of the Brendon Formation whereas in the east both high and low-angle faults cause juxtaposition with the Kate Brook Slate. The southern margin is defined by several steep E-W trending faults that produce bold ridges around Pillaton (Fig. l).

The formation is predominantly inverted and the upper and lower boundaries are inferred thrusts. Only at Park Wood is the lower contact exposed and there a flat-lying, brecciated fault zone separates Crocadon sandstones from the underlying black cherts of the Brendon Formation. Elsewhere the basal thrust is marked by a prominent break in slope, changes in surface brash and numerous blocks of vein quartz. The upper thrust superimposes thin sequences of intensely deformed chert and the maximum structural thickness of the formation exceeds 100 m.

Lithology

The Crocadon sandstones are notable for their diversity, ranging from feldspathic greywackes to polymict conglomerates with rapid lateral and vertical variations in grain size. They contain a high proportion of strained and partially recrystallised quartz grains, fresh to altered plagioclase feldspar, muscovite laths and lithic fragments set in a sericitic matrix. The coarsest sandstones reveal phenoclasts of shale, chert, igneous fragments and armoured clay flakes whilst reworked pebbles of chert and sandstone occur within individual beds (Fig. 2.1). Weathering of the fine-grained matrix promotes an open texture and a pale coloration typifies most of the thicker sandstones.

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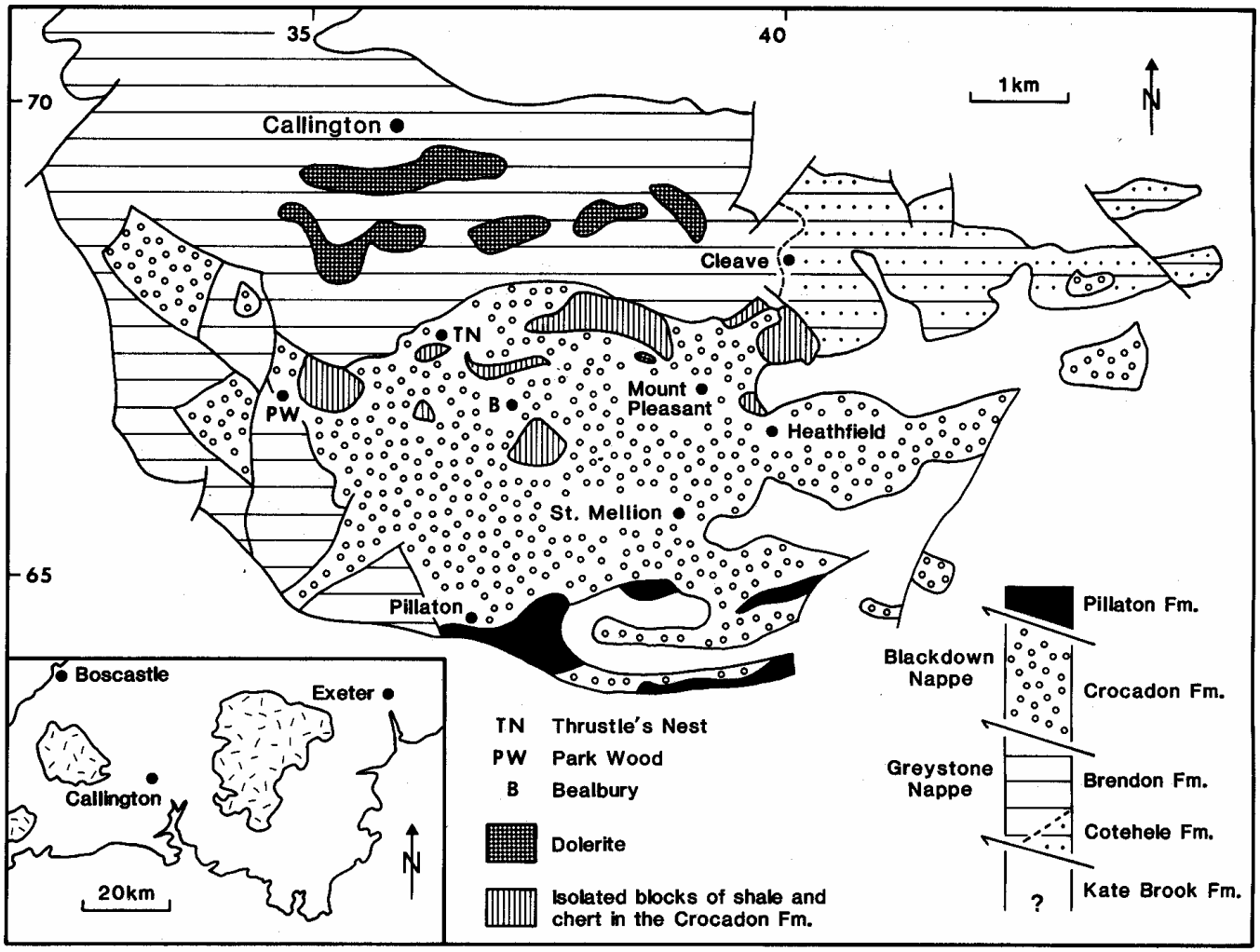

Figure 1. Simplified geological map of the St. Mellion outlier.

The siltstones possess a fine lamination which results from concentrations of well sorted quartz grains and flakes of muscovite. This is occasionally enhanced by opaque ferruginous minerals and carbonaceous films that aggregate preferentially in the micaceous laminae. The shales are chloritic with varying amounts of mica, corrensite and amorphous carbon and their bedding surfaces are flecked with detrital mica and plant fragments. Comminuted plant debris is conspicuous in all the Crocadon lithologies but at Thrustle's Nest an intact plant horizon has been identified. There, at the interface between a sandstone and shale, a mass of simple fibrous stems penetrate the underlying shale and demonstrate that at least some of the plants were firmly rooted and growing in this sediment.

Sedimentary and deformational structures

Graded bedding is recognisable in the Crocadon Formation but it is more common for the sandstones to show compositional grading with abrupt changes in grain size occurring across scoured surfaces (Fig. 2.2). In contrast the thinner sandstones are better sorted, mainly ungraded and only have faint indications of parallel or

Figure 2.1. Crocadon sandstone with reworked pellets of black shale and chert associated with comminuted plant debris. The small pebbles are about 30 mm long. Temporary excavation on Viverdon Down (SX37406746), St. Mellion, Cornwall.

2.2. Crocadon sandstone with poorly sorted horizon showing a scoured upper surface overlain by fine siltstones exhibiting weak crosslamination and small flame structures. The top of the bed as found in the field is arrowed, indicating that the unit is inverted. The specimen is 155 mm long. Temporary excavation on Viverdon Down (SX37866730), St. Mellion, Cornwall.

2.3. Cylindrical and elliptical sand-inflllled burrows resembling Planolites preserved in grey siltstone. Mean burrow diameter is 5mm. Temporary excavation on golf course (SX37846527), St. Mellion, Cornwall.

2.4. Transverse section showing two apertures of a U-tube burrow assigned to Diplocratenion. The dumb-bell-shape outline is produced by a positional shift of the connecting bend during burrowing. Diameter of each burrow is about 15 mm. Temporary excavation on golf course (SX37826527), St. Mellion, Cornwall.

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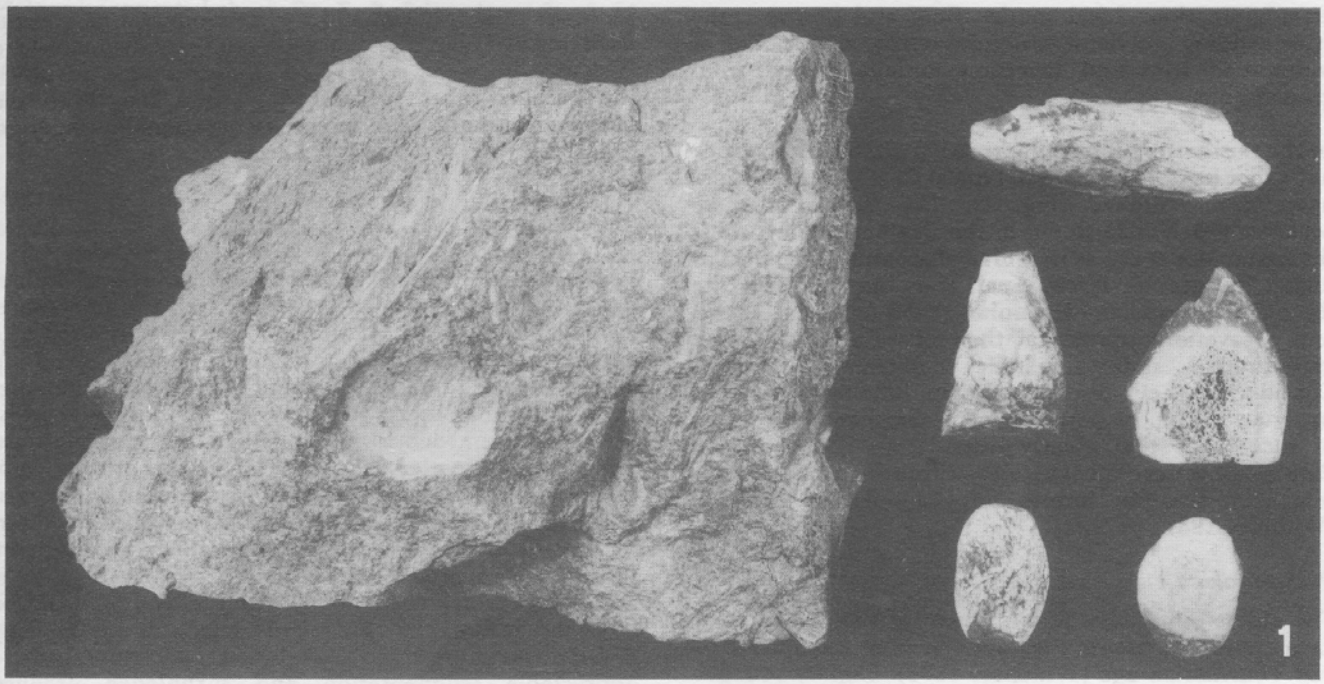

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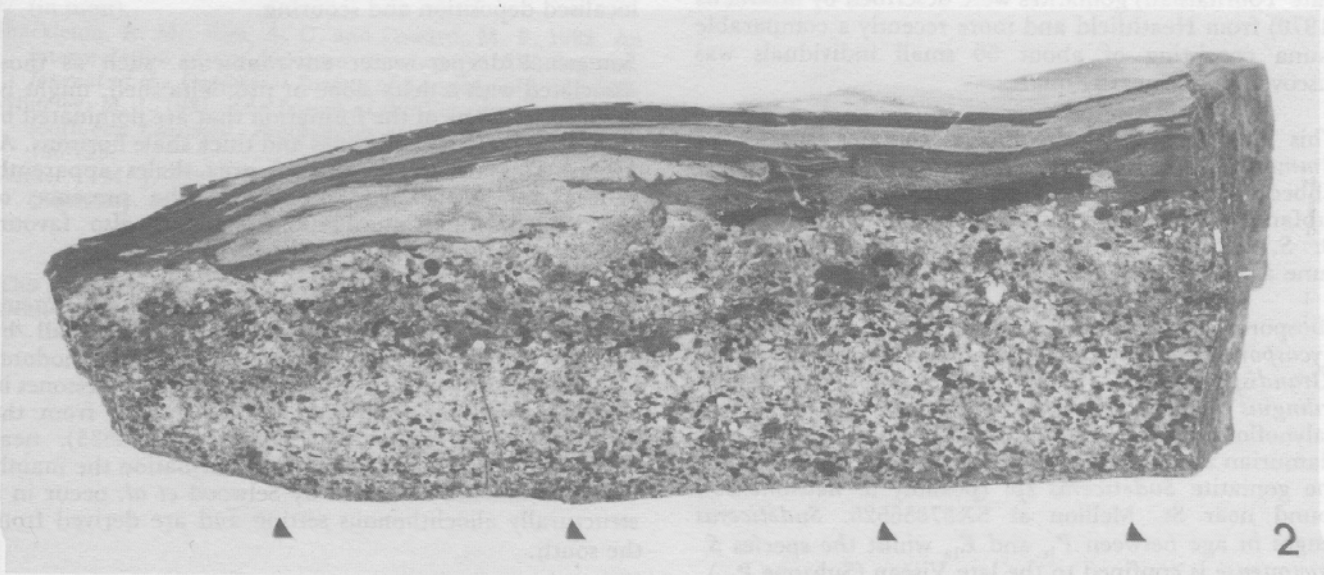

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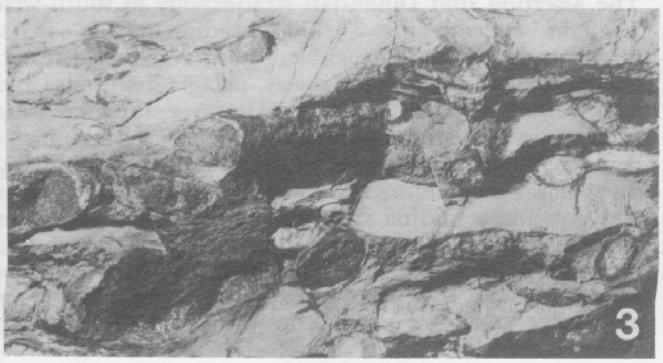

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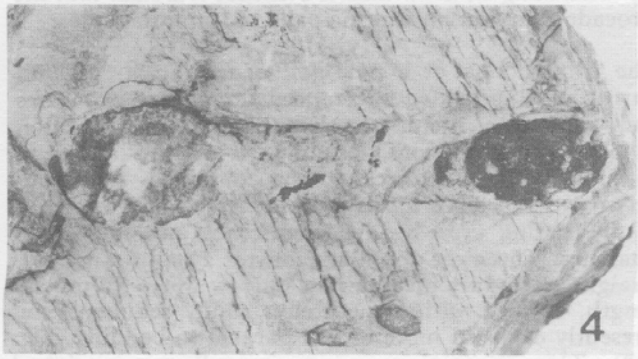

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cross-lamination. Load marks, flute and groove casts, cross-bedding and flame structures reinforce the notion of widespread structural inversion initially gained from graded units.

In the northern half of the formation ten individual rock units that bear no relation to typical Crocadon lithologies have been delimited (Fig. 1). The largest of these occupies nearly 0.25 sq km in the Bealbury valley whilst the remainder occur as irregular, discontinuous blocks at higher topographic levels. Apart from one small dolerite the blocks are either Famennian shales or Viséan cherts and excavations prove that their boundaries are never gradational and sometimes obviously faulted. Considering these discrepancies in lithology, age and structure the blocks are regarded as fragmented elements of successions that clearly differ from the Crocadon Formation yet are incorporated within it, probably as a result of tectonic imbrication caused by thrusting (Whiteley 1981).

Palaeontology

The oldest indigenous fauna presently known from the formation is the Mount Pleasant conodont assemblage of mid-Tournaisian age (Whiteley, 1981). Slightly younger (late Tournaisian) goniatites were described by Matthews (1970) from Heathfield and more recently a comparable fauna consisting of about 50 small individuals was discovered in a nearby quarry.

This new goniatite assemblage contains species of *Ammonellipsites* and *Muensteroceras* plus two strongly ribbed, moderately involute forms referred to the subfamily Karagandoceratinae. In the opinion of the late Dr. S. C. Matthews both faunas can be regarded as the same age.

Miospore floras are dominated by long-ranging *Lycospora pusilla* but associated forms include *?Grandispora sp., Punctatispontes spp., Tripartites trilinguis* and *Triquitrites cf. marginatus.* This palynoflora is indicative of the late Viséan-early Namurian and further evidence of that age is provided by the goniatite *Sudeticeras sp.* (possibly *S. newtonense*), found near St. Mellion at SX37856525. Sudeticeras ranges in age between P1c and E1a whilst the species S. newtonense is confined to the late Viséan (Subzone P_{2b}). By combining all the available evidence it is clear that the Crocadon Formation is principally of Dinantian age.

The construction of a new golf course at St. Mellion during 1983 provided substantial temporary exposures and allowed a more complete analysis of Crocadon lithologies than previously possible. At an horizon close to that in which *Sudeticeras* was obtained numerous occurrences of trace fossils were identified in a sequence of alternating cross-bedded sandstones and micaceous siltstones. The most abundant form is a non-branching, straight to slightly meandering burrow up to 20 mm in length, infilled with structureless sandy sediment and presently disposed more or less parallel to bedding (Fig. 2.3). This orientation appears to reflect an original horizontal burrow of the *Planolites* type although it is possible that some specimens initially formed as vertical burrows (cf. *Skolithos*) and have been re-aligned by processes of late burial and the development of a weak, flat-lying cleavage, less common but more distinctive are U-shaped burrows, referred to *Diplocraterion*, in which the parallel tubes are up to 70 mm long with the interconnecting septum or spreite occasionally preserved (Fig. 2.4).

Discussion

The diversity within the Crocadon sequences is best explained in terms of comparatively shallow-water sedimentation, much of which could be attributed to a deltaic environment, Common features of a delta-top region include rapid lateral changes in bed thickness and lithology, pronounced stratification of individual beds, abundant plant debris and cross-bedded channel sands. In addition, rootlet beds and nodular horizons form during the periodic establishment of subaerial conditions. All but extensive, channel-fill sandstones have been recognised in the Crocadon Formation. Further support for shallow-water conditions is provided by the trace fossil assemblage which corresponds with the *Skolithos* and/or *Cruziana* ichnofacies (Frey and Seilacher, 1980); both develop on littoral to infralittoral substrates where appreciable sedimentation and turbulence produce localised deposition and scouring.

Somewhat deeper water environments, such as those associated with a delta slope or pro-delta shelf, might be invoked for parts of the formation that are dominated by thin, structureless sandstones and thick shale horizons. At Mount Pleasant the laminated grey shales apparently reflect low energy conditions and the presence of siphonodellid and gnathodid conodonts also favours increased water depths (Austin, 1976).

The recognition of shallow-water Dinantian sediments has been extended beyond south-east Cornwall by Selwood *et al.* (in press). They identify littoral conodonts and trace fossils, shelly faunas and bioclastic limestones in several formations that map southeastwards from the north Cornwall coast to Trenault (SX264833), near Bodmin Moor. Like the Crocadon Formation the mainly clastic sediments described by Selwood *et al.* occur in a structurally allochthonous setting and are derived from the south.

The assertion that Dinantian stratigraphy in south-west England reflects impoverished sedimentation in a relatively deep and non-turbulent basin remains justified; of that the abundant black shales, pelagic limestones and radiolarian cherts are testimony. However, the increasing evidence of a contemporaneous shallow clastic facies demands some modification of the overall view, implying that a southern landmass was actively supplying sediment and creating a marine shelf during Dinantian times. In order to establish the nature and location of that source region detailed sedimentological studies of the Crocadon sandstones and their correlatives are required and some constraint mint be placed on the scale of northerly directed translation.

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Acknowledgements I thank Dr. E. B. Selwood and Dr. J. M. Thomas for their sustained interest and guidance, Dr. R. Goldring for helpful discussion concerning the trace fossils and Dr. R. N. Riley for identifying and commenting on the age of *Sudeticeras* from St. Mellion.

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Read at the Annual Conference of the Ussher Society, January 1984

Compositional variation in the lithium micas. (Abstract).

JOANNA S. MARTIN

Joanna s. Martin, 36 Mansfield Close, Oakwood, Warrington, Cheshire.

The St. Austell granite in Cornwall contains unusually high concentrations of Li_2O. The majority of the Li is concentrated in the Li-micas.

The chemical compositions of over 200 individual grains of Li-mica from the megacrystic and non-megacrystic Li-mica granites were determined and studied. Analyses were obtained using the electron microprobe for Si, Al, Ti, Fe, Mn, Mg, Ca, Na, K, F, Rb and Cs. Li determinations were made with the ion microprobe using a technique known as secondary ion mass spectrometry (SIMS).

Using Foster's (1960) classification of the Li-micas, the micas analysed were shown to range from Li-free muscovite from the hydrothermally altered Fluorite granite, to lepidolites from the fresh non-megacrystic Li-mica granite. Zinnwaldites and siderophyllites were also identified.

An examination of Li_2O, FeO and F zoning in individual mica flakes showed different zoning patterns emerging in micas from the megacrystic and non-megacrystic Li-mica granites. The former showed FeO concentrations increasing towards the centre of grains, whilst Li_2O concentrations increased towards the margins of the grains. The latter showed varying trends for both Li_2O and FeO. Micas from both granite types exhibited F concentrations increasing towards the margins of the grains.

An attempt was made to identify possible end member substitutions taking place in micas from the two granite types. It was shown that micas from the non-megacrystic Li-mica granite contain Li replacing both octahedral Al and Fe^{2+}, the substitutions taking place are:

$$Al^{3+} = 3Li^{+}$$
$$\text{and} \quad Fe^{2+} = 2Li^{+}$$

These trends are not shown by micas from the megacrystic Li-mica granite.

Li and F molecular proportions in micas from the megacrystic Li-mica granite have a positive relationship, however this link is not extended to micas from the non-megacrystic Li-mica granite. Micas from both granite types show the same relationship between molecular proportions of Fe and F, that is a positive correlation at low concentrations of F and a negative correlation at high concentrations of F.

Whilst the results outlined above may not be used to define the mode of formation of the two granite types, they do show that the micas, and therefore the granites are substantially different. There is little doubt that the non-megacrystic Li-mica granite is essentially a late stage magmatic differentiate. The origin of the megacrystic Li-mica granite is a more controversial topic. However the results of this study do not preclude the theory that it may be a metasomatic alteration product of the adjacent biotite granite.

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Note on an Early Devonian fossil locality on the Dart Estuary, south Devon.

K. M. EVANS

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In the course of mapping the Torquay area (Sheet 350), the Geological Survey noted the existence of a fossiliferous locality on the east bank of the River Dart, near Lower Noss Point. The beds here were mapped as Meadfoot Beds and yielded brachiopods referred to *Spirifera primaeva* or *Spirifera Decheni* by Ussher (1903), subsequently redetermined as *Spirifer cf. fallax* by Lloyd (1933).

During an investigation of the Lower Devonian brachiopod faunas of south-west England (Evans, 1980a), the writer was able to examine the Geological Survey material and also make new collections from this locality i.e. south of Lower Noss Point where fossil remains occur in friable brown and yellow shales, immediately south of wooden piles in river bed (Grid reference SX88005265).

The remains are very poorly preserved, usually in low relief with varying amounts of distortion, however, bulk collections enabled the following fauna to be identified: *Rhenostrophia cf. gigas, Oligoptycherhynchus daleidensis, Acrospirifer primaevus* and *Hysterolites hystericus* in addition to solitary and compound corals, gastropods and crinoid fragments. The brachiopods are indicative of a mid or late Siegenian age.

This locality is of interest when considered in relation to the Meadfoot Group (Harwood, 1976) as a whole. The faunal and lithological characters of this locality compare closely with the Meadfoot facies (Evans, 1981) of the Meadfoot Group of Cornwall which has yielded brachiopods indicative of a similar age (Evans, 1980a; 1980b), in particular, the locality of Polyne, near Looe. Polyne Quarry has long been recognised as an important locality. (Green and Sherborn, 1906; Ussher, 1907) and recent collections have yielded a late Siegenian brachiopod fauna (Evans, 1980a) from beds very similar to those cropping out in the Dart estuary.

The River Dart locality does not resemble the Meadfoot Group around Torquay which is entirely Emsian in age (House and others, 1977, Evans, 1980a) and does not appear to be comparable with the poorly fossiliferous outcrop of the Meadfoot Group in the South Hams district (Ussher, 1904).

Although a detailed investigation of the area is needed, it would appear that the exposure on the River Dart could represent an allochthonous block, derived from the west. Such an occurrence might be explained by the effects of thrusting, which has been described from this part of south Devon (Coward and McClay, 1983). This is only a tentative suggestion and considerable fieldwork and palaeontological investigation is still required, however, it is hoped that this observation may serve to stimulate research in this area.

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British Triassic palaeontology: supplement 8

G. WARRINGTON

British Geological Survey, Keyworth, Nottingham NG12 5GG

Since the submission of the writer's previous supplement (Proceedings of the Ussher Society, 5, 493; 1983) to his paper on British Triassic palaeontology, the following works dealing with or including aspects of that subject have appeared:

Ambrose, K. 1983. Sheets SP06 and SO96 (eastern part) (Redditch and Feckenham). Quaternary deposits with special emphasis on potential resources of sand and gravel. *Geological Report for DoE; Land Use Planning*: Institute of Geological Sciences, Keyworth, 26pp.

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Information for Authors

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4. Manuscripts should be typewritten (preferably on A4 size paper) on one side only, and double-spaced throughout (including references, figure captions, tables, etc.), with ample margins; **two copies must reach the Editor in a complete state not later than 1st February.**
5. Papers should be laid out as in the current issue of the *Proceedings*.
6. Illustrations must be clearly marked with the author's name and figure number. Plan figures so that they take up the entire width of the type area (157mm), or the width of one column (76mm). When preparing a full page illustration, allow space for the caption to come within the page depth, 205mm. Prepare illustrations at 1 1/2 times final size. Photographs should be printed on glossy white paper, and must be clear and sharply contrasted, but without pronounced light areas and heavy shadow. Mount photographs on stiff white card.

Notes and Abstracts

1. Notes are short communications giving the results of work in progress, accounts of techniques, and other brief accounts of matters not normally suitable for treatment as a full paper. Notes will generally be submitted to referees. The length of a Note will be considered mainly in relation to its content.
2. An abstract should be a brief summary of the contents of the paper read at the Annual Conference and should not normally exceed 300 words in length. Text figures and references are normally not allowed. Abstracts are not submitted to referees and will be published with minor editorial corrections only.

A fuller version of the Society's "Preparation of manuscripts for the *Proceedings of the Ussher Society" is* available on application to the Secretary. Manuscripts and enquiries on editorial matters should be directed to the Editor at the following address:

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