# Proceedings of the Ussher Society

Edited by G.M Power

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# Passive margin subsidence: evidence from Rockall Plateau and the Western Approaches (Abstract)

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Passive margins represent the broken edges of continental plates which have been moved apart by sea floor spreading. One of their attributes is that they have undergone major subsidence. Various factors influence the rate and amount of subsidence and among these may be listed initial thermal doming or crustal thinning with an enhanced geothermal gradient, thermal cooling with age and distance from the spreading centre, and loading due to accumulation of sediment. In order to test geophysical models of such subsidence, it is essential to be able to determine the depth of deposition of passive margin sediments. Both the Western Approaches and the south-west margin of Rockall Plateau are passive margins on which the sediment successions are thin (starved) and this has permitted drilling down to basement.

On Leg 48 of the Deep Sea Drilling Project site 401 was drilled in 2555m of water in the Western Approaches. It penetrated 341m into the sea floor through a succession of Quaternary to early Cretaceous oozes and chalks beneath which there are about 70m of limestones. The latter are of reefal type with calcareous algae and are of late Jurassic to early Cretaceous age. They were clearly laid down in very shallow water, in an epicontinental sea which stretched across Europe and the adjacent Grand Banks area of Canada. During the time of their deposition rifting was taking place and considerable sea floor relief developed. In the early Cretaceous (Aptian) the two continental plates separated and sea floor spreading commenced. However, no depth determinations of the Cretaceous deposits have been published, partly because it is difficult to determine the depth of water with precision from Mesozoic faunas.

Rockall Plateau is a microcontinent; sea floor spreading commenced between it and Greenland in the early Eocene, following a period of rifting in the Palaeocene. Two sites were drilled on Leg 48 of the Deep Sea Drilling Project and a further 4 sites on Leg 81 (on which the speaker served as a shipboard scientist). Early Eocene clastic sediments were found to overlie basaltic lava flows of late Palaeocene age. These Eocene sediments contain plant debris and macrofauna such as oysters and *Nucula.* Their foraminiferal faunas are similar to those of Europe. The planktonic: benthic ratio is low, indicating isolation from the open ocean. The benthic foraminifera have a low to medium diversity comparable with that of modern brackish lagoons or estuaries and marine continental shelf in a temperate region. The abundance and nature of individual genera and species supports these environmental interpretations. The middle and late Eocene successions are condensed and incomplete. The supply of clastic detritus was cut off (through the subsidence below sea level of Rockall Plateau). The foraminiferal faunas indicate a deepening of the water to outer shelf and upper slope depths. The Oligocene is deeper still and the Neogene was deposited in depths not unlike those in the area at present (2.2 - 2.5km). When plotted against time, the amount of subsidence determined from the palaeoecological interpretation closely agrees with that calculated from geophysical theory.

Other points of interest that arise from these observations are, first, that the sedimentary successions are incomplete with numerous, often prolonged hiatus breaks, and second, that lithology is not a reliable guide to the depth of deposition or the existence of hiatuses.

It is hoped that this resumé of subsidence on Mesozoic and Cenozoic passive margins will stimulate thought on the possible role of such processes in the Palaeozoic history of south-west England.

# Devonian eustatic events

M.R. HOUSE

House, M.R., 1983. Devonian eustatic events. *Proc. Ussher Soc. 5*, 396-405.
The pattern of major and minor sedimentary rhythms and cycles of the New York Devonian is taken as a starting-point for a discussion of the recognition of international Devonian eustatic events. Attention is drawn to the consistency of evidence for eustasy in the Frasnian clastic rhythms of New York and the reef complexes of Western Canada and Belgium. Models are proposed for more detailed sedimentological and biostratigraphical scrutiny.
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## Introduction

Little has been written on international events in the Devonian which may reflect global sea-level changes. On the other hand this subject has given rise to a substantial literature in other parts of the Phanerozoic having been considered generally in relation to seismic stratigraphy by Vail *et al.* (1977, 1981) for the Mesozoic and Tertiary, and more specifically for the Cretaceous by Hancock and Kauffman (1979), for the Jurassic by Hallam (1978)and Wilson (1980), for the Carboniferous by Ramsbottom (1979) and for the Lower Palaeozoic by Legget *et al.* (1981). Details relating to Devonian transgressions around the Old Red Sandstone continent have, however. been given by the writer (1975) and there is a relevant slim booklet of abstracts (Embry *et al.* 1982) on controls on sedimentation of the American Devonian.

In this paper it is intended to extend the earlier review with rather fuller biostratigraphical cover and to outline areas where detailed analysis is desirable and to comment on methods of quantifying Devonian eustatic events. The New York Devonian is taken as a starting-point. In such studies the aim is to attempt to elucidate one of the three independent variables of sedimentation, namely, local subsidence rate, sedimentation rate, and eustatic change of sea level. The first step is the assembly of detailed evidence from individual areas with refined time documentation. The second step is the comparison of evidence in different regions.

The area normally the concern of the Ussher Society includes Dorset where a pattern not untypical of shelf sedimentation is seen (Fig. 1). The rhythmicity of the succession here was first remarked on by Conybeare and Phillips in 1822. Subsequent work is summarised in a review and discussion by Wilson (1980). Six such rhythms for the Dorset Jurassic are indicated on Figure 1. Arkell suggested ten for the Jurassic of the South of England in 1933 but he included three rhythms within the Corallian Beds (revised by Talbot 1973) and two between the Fuller's Earth and Cornbrash and he hinted at, but did
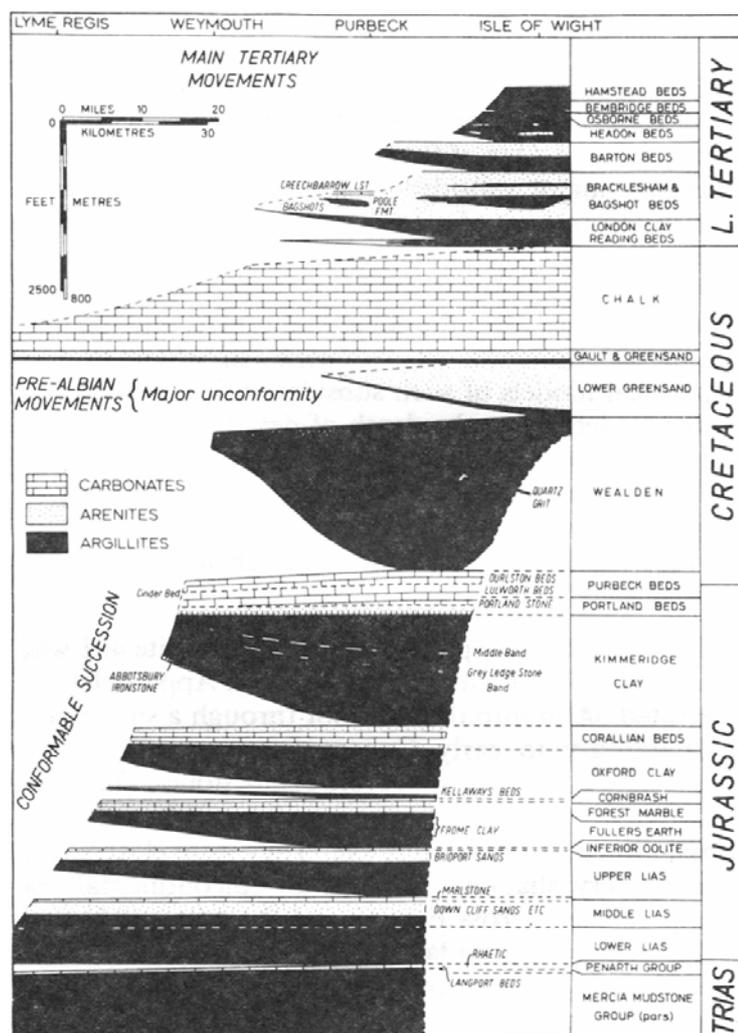


Figure 1. Diagram of the Mesozoic and Tertiary sequence of Dorset and the Isle of Wight drawn to emphasise rhythmicity. Jurassic arenites are mostly fine-grained and often dolomitic siltstones.

not elaborate on, rhythms within the Portland Beds which have since been described by Townson (1975). These give a gradation in scale from total thicknesses of 10-15m (for example the Corallian Nothe Clay/Bencliff Grit rhythm up to at least 580m (the Coastal Kimmeridge Clay/Portland Beds rhythm). Clear diastems not having been demonstrated for the most part in shore-ward facies the synthem terminology of Chang (1975) is not yet appropriate.

At yet finer scales are the smaller rhythms of the Blue Lias (Sellwood *et al.* 1970), Bathonian (Penn *et al.* 1979) or Kimmeridge Clay (reviewed in Wilson 1980) where the repeated rhythms may be only a few metres thick. At still finer scale are the microrhythms which have been interpreted as annual varves.

Such gradation in scale varies with position in the depositional basin so that clear terminology is not simple. Terms such as macrorhythm (over 10m), mesorhythm (1-10m), and microrhythm (up to 1m) might be useful, but in the succeeding account of the New York sequence the maximum thicknesses of outcrop known for a rhythm will be quoted in parenthesis where possible. What is of interest in the Jurassic example is the way in which the facies of the macrorhythms are repeated in the mesorhythms and how both suggest shallowing upward and anoxic to oxic environmental change. This too is the broad pattern of the Devonian examples.

## New York Devonian

It has long been recognised that the New York Devonian is one of the best, if not the best, development of the system in the world (Fig. 2). The rhythmic pattern of its major and minor sedimentary units has also long been recognised, first for some of the transgressive black shale tongues by Hall in 1843, and later some of these rhythms have formed the basis for a formational and group terminology (Rickard 1975). The setting shows coarse clastics of the Catskill Delta complex to the east, especially in the middle and late Devonian with units thinning westward and craton-wards where, curiously, intra-cratonic basins were the site of largely black shale

Figure 2. The New York Devonian showing major divisions and their relation to deeper water facies tongues passing eastward into the Catskill Delta terrigenous facies (modified from Rickard 1964 and House 1975).
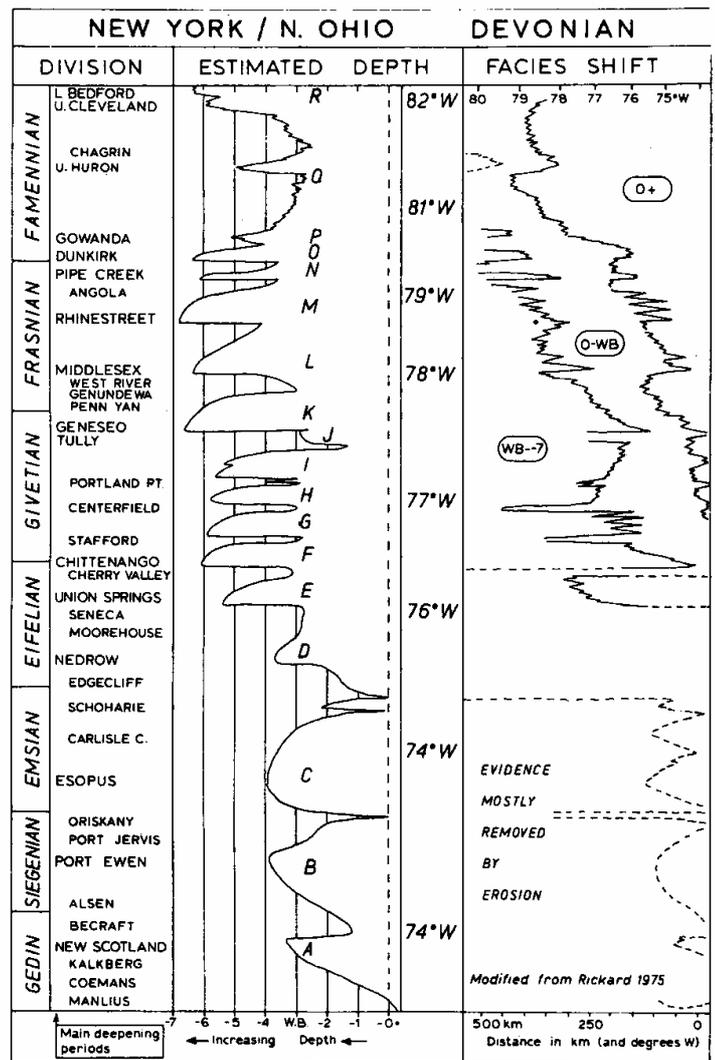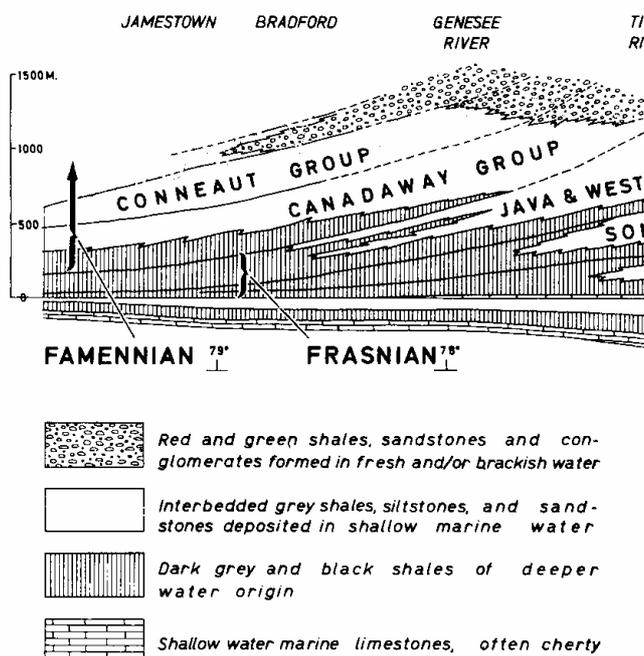




Figure 3. Diagram summating evidence on depth changes and facies shifts for the New York Devonian. Data from many sources but facies shift data from Rickard 1975. (0, approxin sea level. W.B., approximate wave base).

deposition in Ohio and adjacent states especially in the late Devonian. Limestones decrease in importance upwards from the Lower Devonian Helderberg Group, and early Middle Devonian Onondaga Limestone, and a yet more limited Tully Limestone higher. Limestones in the Upper Devonian (Fig. 4) are virtually absent. 'Thus facies transects and successions differ markedly from those well described from Europe where, in the south west England, Ardermes and Schiefergebirge, the Lower Devonian is arenitic, the Middle largely carbonate and the Upper argillitic to arenitic in the marine facies.

An attempt has been made to summarise both the facies shifts and the depth of water for the whole New York Devonian in Figure 3. The area deserves far more rigorous quantitative analysis of sediment type, depth of formation, palaeoecological regime, and facies shift than is currently possible, but significant contributions towards this end are being made. Summaries of broad data are given by Rickard (1975) and Oliver *et al.* (1969).

If the period from early Givetian to the end of the

Devonian is taken as 20Ma (as in Harland *et al.* 1982) and since the maximum thickness of shallow water or terrigenous clastics is 3550m for this span, then assuming compaction occurs in early diagenesis and can be

ignored, then the basin subsidence rate is 0.17mm/a

the New Scotland limestone and shales of subtidal facies and a high diversity shelly fauna (maximum thickness about 70m).

**B**. In the Upper Helderberg Group a similar rhythm is repeated commencing with less shallow facies than the Manlius but terminating in the Port Ewen limestones and shales which are similar in type to the New Scotland
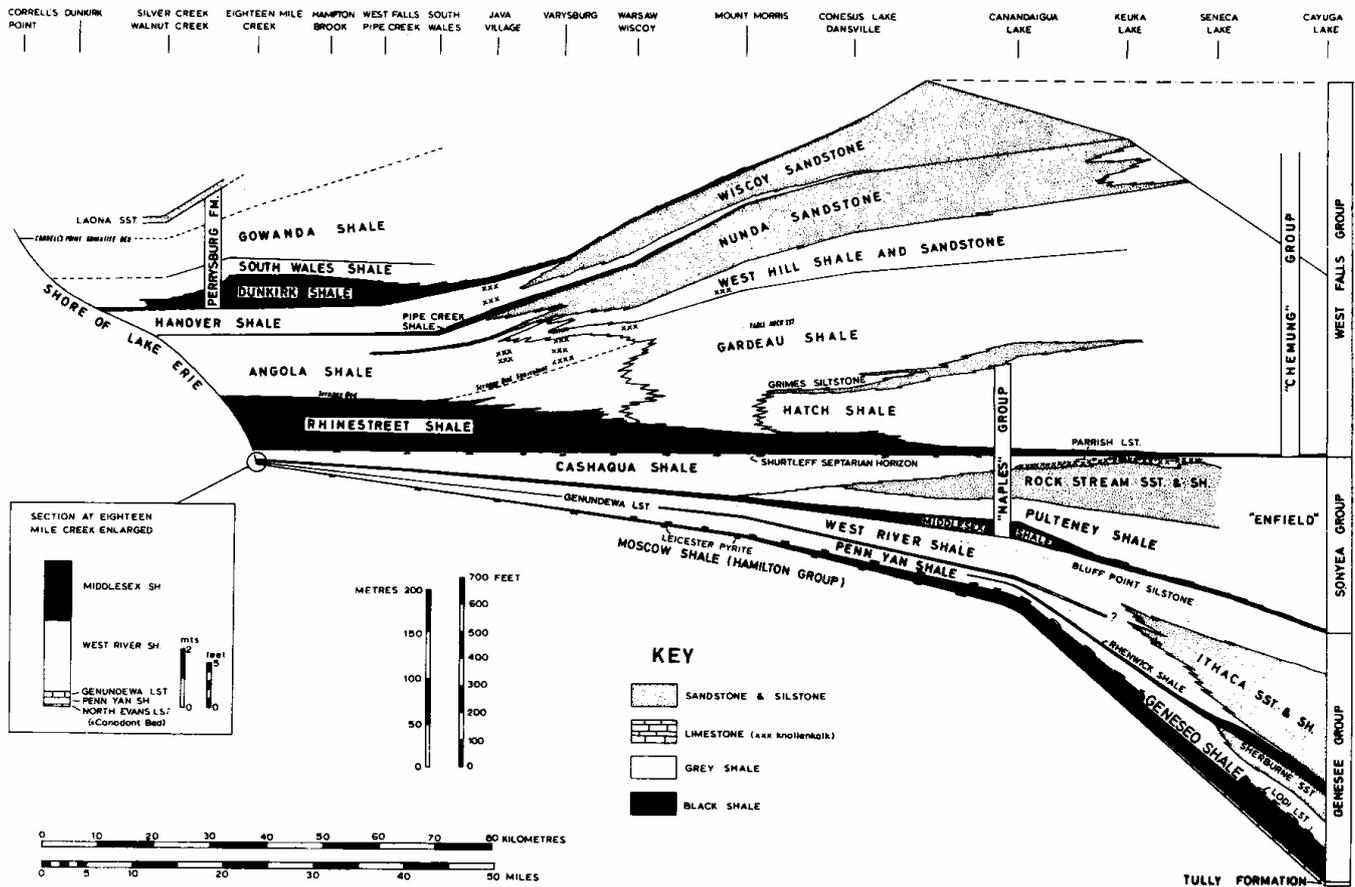


Figure 4. Diagram of Facies changes in the Frasnian and adjacent rocks in New York. From Kirchgasser and House 1981

between longitudes 74° to 76°W, and perhaps an order more than that in the intra-cratonic black shale basis to the west.

There is an overall facies shift during the Devonian (Fig. 2) which results in the change in type of the rhythms described below through time. The ideal of studying such changes only within one facies is not possible.

*Lower Devonian*
Data for the Lower Devonian is mostly contained in works by Rickard (1962, 1975) but studies of small scale rhythmicity are commencing (Anderson and Wilson in Embry *et al.* 1982).

**A**. The earliest rhythm in the Lower Helderberg Group shows progressive deepening and starts with calcilutites and dololutites of supratidal type in the lower Manlius and passes upward through biostrome and bioherm facies to

(Maximum thickness about 65m).

**C.** Overlying the Helderberg Group the geographically restricted Port Jervis limestones pass up into calcilutites which are laterally the equivalents of the high energy subtidal Oriskany arenites and suggest a shallowing upward rhythm probably terminated by erosion.

**D**. The remaining units of the Tristates Group comprise the Esopus Shale of siliceous shales and cherts with rare goniatites which grades up into the calcareous shales and siltstones of the Carlisle Center and culminates in the variable argillaceous limestones of the Schoharie with a rich shelly fauna. The whole suggests a sequence of increasing shallowing and oxicity upwards, (Maximum thickness of 3 and 4 about 229m).

The general facies and palaeoecology and community types for the Appalachian Lower Devonian have been

reviewed by Boucot and Johnson (1967) but the detailed record of facies and community shift has not been documented. As is indicated on Figure 2 cyclicity is indicated as well as rhythmicity.

*Middle Devonian*
The Middle Devonian of New York comprises the essentially carbonate Onondaga Formation, together with the overlying rhythmic Hamilton Group with periodic thin limestones. The transgressive Tully limestone, representing the Taghanic onlap of Johnson (1970), is now recommended for inclusion in the Middle Devonian by the Devonian Subcommission.

**E**. The Onondaga Formation represents a major transgressive event in eastern North America and it carries carbonate facies westward over cratonic areas, a distribution reviewed by Oliver *et al.* (1969). Detailed analysis of Onondaga facies is given by Oliver (1954, 1956) and Koch (1981). Early bioherms (Edgecliff)give way to deeper chert and shale facies (Nedrow) and the facies shallows up into coral and braciopod carbonates (Moorehouse and Seneca) in central New York. The Edgecliff is interpreted as belonging to the *patulus* Zone s.1. (Klapper 1981), now divided into *patulus* (s.s.)and *partitus* Zones with the division between them the boundary recommended to define the base of the Middle Devonian by the Devonian Subcommission (Fig, 7). Erosion preceded the Onondaga deepening and a cyclic, rather than rhythmic pattern is indicated. Finer scale microrhythms are described within the Onondaga by Oliver (1956). (Maximum thickness 76m).

**F**. A rhythm is initiated with the black shales of the Union Springs which terminates with the shallower carbonates of the Cherry Valley Limestone which succeeds the Seneca or may, in part, be equivalent to it. The Werneroceras Bed at the top of the Union Springs contains *Cabrieroceras* and can be traced in the southern Appalachians (House 1978). Klapper refers the Werneroceras Bed and Cherry Valley Limestone to the *kochelianus* Zone (Fig. 7) and hence deduces these beds are top Eifelian in European usage. (Maximum thickness ca. 12m or more).

**G**. Subdivision of the rhythmic and cyclic units of the late Hamilton is somewhat subjective, but the broad pattern was clearly elucidated by Cooper (1930) and he also recognised smaller scale repetitions. More detailed discussion has been given by McCave (1973), Brett and Baird (1982) and many others.

An initial Hamilton rhythm is indicated by the deeper black shale facies of the Chittenango which pass up to the shallow water facies, rich in benthos, culminating in the Stafford Limestone. The nature of the actual facies shifts changing from west to east as described by Cooper with the entry of the siltier Cardiff Shale bearing a shelly fauna. (Maximum thickness over 171m).

**H**. The Levanna Shale introduces black shales in the west which grade up to the Centerfield Limestone but with an intervening Pompey black shale. Cooper *et al.* (1942) identified the Centerfield level as far west as Arkansas, a more extensive distribution than for any other Hamilton limestone horizon. (Maximum thickness at least 75m).

**I**. The complexities of the New York Upper Hamilton are more fully realised as a result of the work of Baird (1979). The black shales of the Ledyard introduce deeper water facies, again with *Leiorhynchus,* above the Centerfield but higher there are several calcareous levels, including the Tichenor, Menteth and Portland Point limestones, the last two defining the base of the Moscow Formation. McCave (1973) attempted to quantify sea-level changes associated with a Portland Point transgression which he regards as a temporary event in the regional regression. Several sea-level changes are represented in the Ludlowville and Moscow formations (Maximum 150m or more). The Centrefield appears to be L. *varcus* Zone in age (Klapper 1981) and the upper Mamilton above appears to be early *M. varcus* Zone.

**J**. The major transgressive unit, the Tully Limestone, has been recognised as one of the two major events of the American Devonian for a century and has been well documented (Heckel 1973) and the transgression reviewed (Johnson 1970, House 1975), the event falling within the later M. *varcus* Zone and extending into the U. *varcus* Zone. What is new is the recognition that much record has been removed by pre-Genesee or pre-Genundewa erosion (Brett and Baird 1982). If considered as separate from the Moscow, maximum thickness is about 100m in eastern Clastic facies.

*Upper Devonian*
The Devonian Subcommission is currently recommending a level for the base of the Upper Devonian which may lie somewhere just below the Genundewa rather than the traditional level below the Tully (House 1982). The new level is as inconvenient in New York as in many places. The sequence around the Frasnian of New York (Figs. 4,5) shows as well as anywhere typical Devonian rhythmicity; some units are strictly rhythms (Fig. 6B), some have cyclic characters (Fig. 6A). The sequence is well established through the work of de Witt and Colton (1959, 1978) and contributions by them and others to the U.S.G.S. Oil and Gas Investigation Charts. A review is given by Rickard (1975). Significant local palaeocological studies have been made in shelly eastern areas, particularly by Sutton *et al.* (1970) and Thayer (1974). This account includes diagrams and comments resulting from twenty years work by Dr W.T. Kirchgasser and the writer (see Kirchgasser and House 1982).
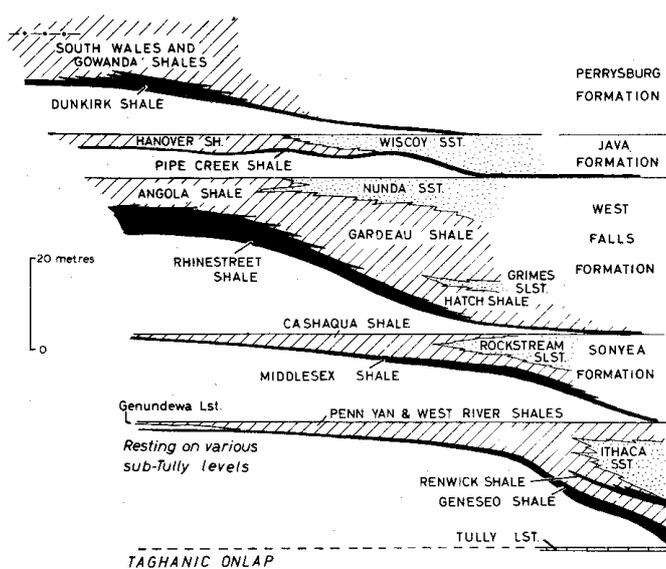
Figure 5. Diagram of facies changes in the Frasnian and adjacent rocks in New York drawn to emphasis rhythmicity.

**K**. The Geneseo black shale indicates a significant deepening event but the base changes date perhaps a result of contemporary bevelling (Brett and Baird 1982). The deepening started in the late U. *varcus* Zone and the Renwick Shale may indicate a post *dengleri* Zone deepening interrupting the progressively shallowing Penn Yan sequence near the top of which *L, asymmetricus* Zone conodonts occur (indicating to the Devonian Subcommission basal Upper Devonian). The succession may be interpreted as indicating maximum regression with the furthest Ithaca Sandstone tongues giving a cyclicity to the upper part. (Maximum thickness 488m).

**L**. The Middlesex black shale and succeeding Cashaqua Shale comprise the Sonyea Formation (or Group) illustrated in Fig. 6A which shows a cyclical return of facies at the top. Klapper (1981) refers the Cashaqua to the *triangularis* Zone. More remains to be done on precise dating, but Sutton *et al.* (1970) give evidence for facies faunas and facies shifts in the areas east of those illustrated here. (Maximum 244m).

**M**. The Rhinestreet Shale marks the largest of the black shale tongues but, as with earlier rhythms, it is composite and comprised of many smaller-scale rhythms up to a few metres in thickness. Some of these in the overlying and interdigitating Lower Angola Shale (Figure 6C) have been traced eastward from Lake Erie for in excess of 100km:

each shows an initiating black shale and the change from anoxic to oxic conditions is marked by a burrowed horizon which is followed by a variable sequence of gray shales usually culminating near the top with a septarian concretion level rich in goniatites. The pattern is similar to the less well documented rhythms of the Geneseo and Cashaqua. If the Rhinestreet transgressive pulse is interpreted as extending a basinal anoxic sapropel facies up over the more easterly prodelta and delta environments then the thickness of the underlying sedimentary wedge, the Sonyea Formation, gives a measure of the deepening involved in the sudden initial Rhinestreet transgression. (Maximum c. 500m). The date may be lowest *gigas* Zone (Klapper 1981).

**N**. The thin black shale pulse of the Pipe Creek Shale (Fig. 4) and overlying Hanover Shale give a rhythm which has been called the Java Group but Rickard (1975) includes it within the West Falls Group. Again the minor rhythmicity continues. In conodont terms a *triangularis* Zone date is suggested (Rickard 1975) and the upper Hanover contains indicators of the goniatite *holzapfeli* Zone so that the event is latest Frasnian in age. (Maximum c. 100m).

**O** and **P**. The prominent black shale of the Dunkirk (O) carries conodonts referred to the lowest Famennian but it is not until the higher Corell's Point Goniatite Bed (Fig. 4) that goniatites of the *curvispina* (or *amblylobum)* Zone of the earliest Cheiloceras Stufe occur; a slight farther deepening (P) is inferred. Higher levels in New York have not been examined in a way which enables either dating or documentation in detail of facies shifts, but the Ellicot Shale suggests a deepening phase probably at the very end of the otherwise regressional Cheiloceras Stufe equivalents.

**Q**. In northern Ohio, as part of the long-continued black shale facies, is an intercalation with pyritic ammonoids referred to the *annulata* Zone at the top of the Platyclymenia Stufe associated with the regressional extension of the Chagrin Shale.

**R**. Higher, in the topmost Cleveland Shale occur pyritised late Famennian ammonoids (House 1978) which suggest a deepening event and a later event is suggested in the lowest Bedford Shale by the occurrence of imitoceratids similar to those of the western Canada Exshaw Formation and the Wocklumeria Stufe of Europe: these two events preceding the more general basal Carboniferous transgression.

**A** MAJOR RHYTHM (SONYEA)  |  **B** MAJOR RHYTHM (L.WEST FALLS)  |  **C** MINOR RHYTHM (ANGOLA)
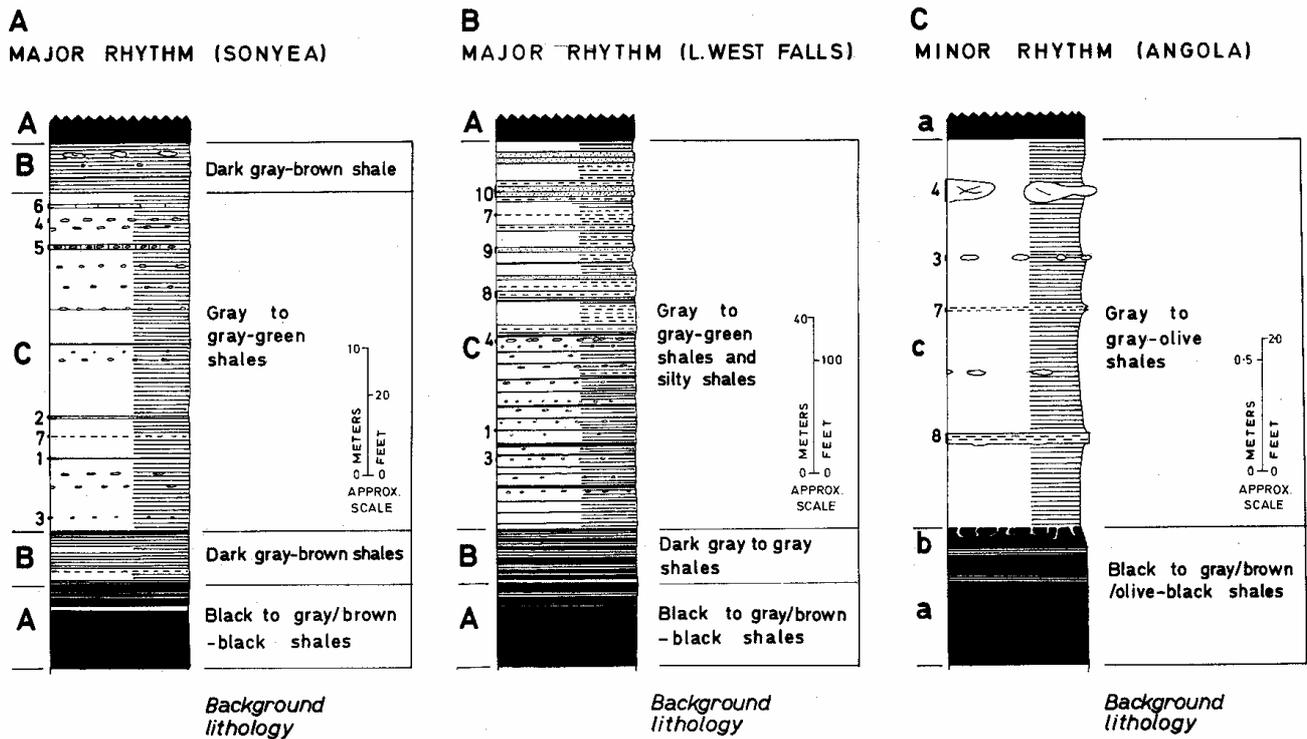
Figure 6. Diagrams illustrating sedimentary cycles and rhythms in the Frasnian of New York. All scales approximate. (1, black shale horizons; 2, bituminous shale; 3, pyritic or calcareous nodules; 4, small to large septarian nodules, sometimes barytic; 5, nodular carbonate unit; 6, carbonate unit; 7, siltstone seam; 8, turbidite unit; 9, siltstone or sandstone unit; 10, massive sandstone unit). A, based on data from W.T. Kirchgasser.

*General Comment*

This succession of essentially deepening rhythms in the New York Devonian enables some plot of deepening phases (Fig. 3) which were superimposed on high general subsidence rate of eastern New York and the record suggesting sympathetic thinning and often penecontemporaneous erosion in western New York which suggests positive response to the easterly subsidence. The scale of depth changes when the movement of the black shale sapropel facies over the clastic wedges (Fig. 5) is considered, suggest a maximum depth for the anoxic black shale layer of at most a few hundred metres, near a figure inferred for the Huron Shale of 150-200m for a transitionally oxic facies by Gutschick and Wuellner (1983). More rigorous calculation of the sedimentary geometry, sedimentation, isostatic expectation, and palaeoecological facies shifts should enable Figure 3 to be considerably refined and quantified.

A test for whether the New York events represent local changes, either of periodic fault controlled local subsidence or major delta movements within the framework of regional subsidence, cannot be answered from New York data. In the next section consideration is given to the extent to which these events can be recognised internationally.

## Review of International Events

Until a large number of detailed stratigraphic sections is documented in many parts of the world and the evidence of facies shifts and deepenings carefully dated the recognition of *international* events will be obscured in detail. Nevertheless the two major transgressive events of the Devonian internationally, that is close to the base of the Middle Devonian and close to the base of the Upper Devonian (the Taghanic onlap) were recognised about a century ago (Fig. 7). Indeed the events were used to define the boundaries (redefinition of the base of the Upper Devonian, alas, has recently obscured this).

Recognition of finer-scale events presses to the limits present biostratigraphic resolution when different facies are involved. For convenience the discussion here uses the New York/Ohio terminology.

*Lower Devonian*
Resolution of international facies shift data following the major effects of the Caledonian Orogeny leave very much to be desired. The New Scotland event (A) may match the deepening leading to the Belgian Mondrepuis Shale following the dating by Bultynck (1982) of the Naux Limestone and similar levels with *remscheidensis* conodonts. The Port Ewen event (B) may subjectively be recognised in Europe (Fig. 7) but there are no convincing facts. J.G. Johnson *(in litt)* has suggested correlation of the McColley Canyon facies shift of Nevada *(kindlei* Zone) with Port Jervis movement (for separated on Fig. 3) and the succeeding regressional Oriskany has some apparently similarly aged shifts in Europe.
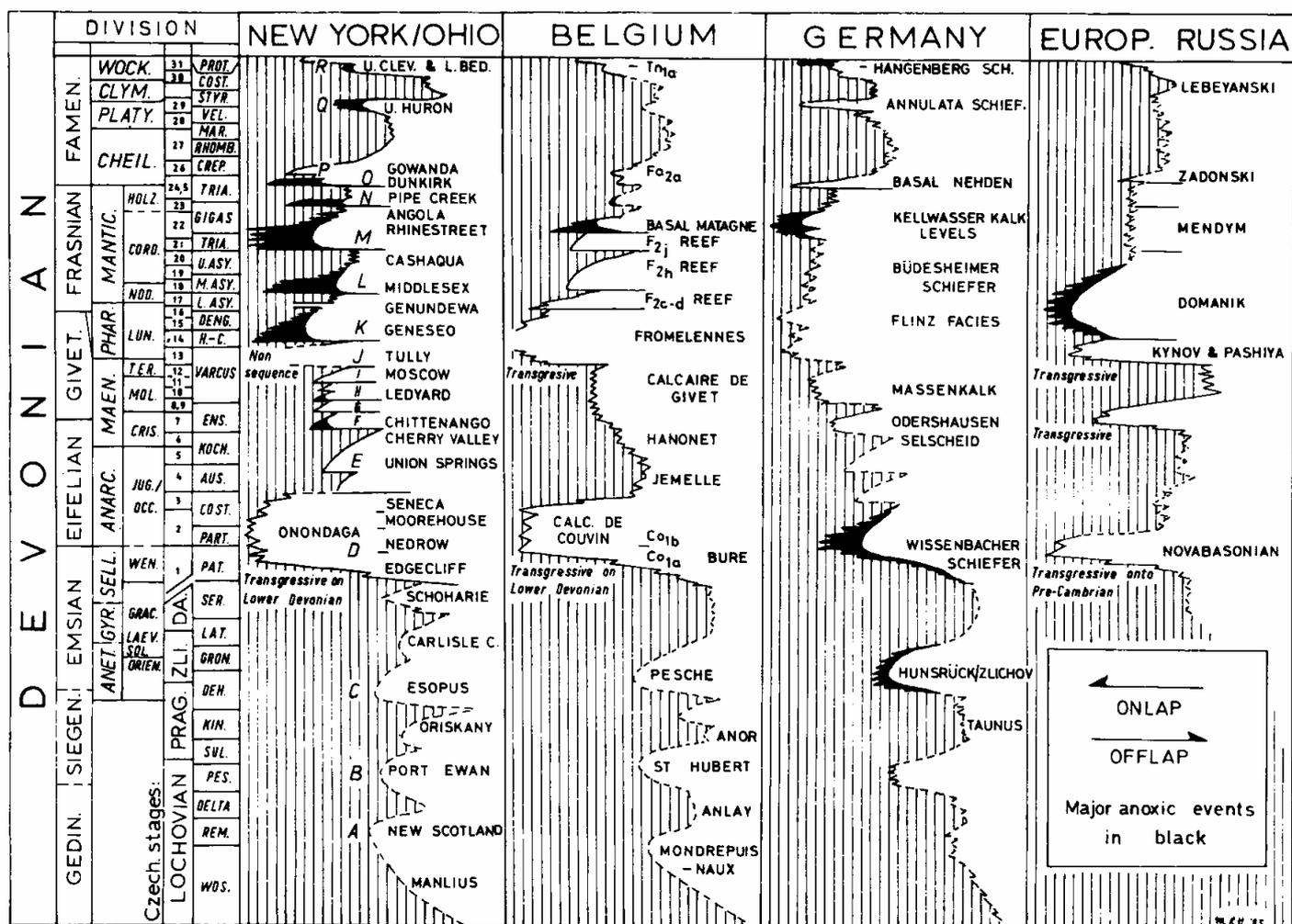
Figure 7. Diagram collating facies movement data for several areas against a Devonian division scale showing stages, ammonoid stufen, ammonoid zones (numbers are based on Appalachian scale of House 1978) and basic conodont zones

The next deepening event, commencing with the Esopus Shale of New York (C) seems to correspond to a widespread deepening event culminating in the Hunsrück Schiefer of Germany, Zlichov to Daleje deepening of Czechoslovakia and the *regularissimus* Zone deepening of the U.S.S.R. but precision in dating is not available. The European data in particular is confused by the marked facies contrasts in different areas and the lack of precise facies shift documentation in individual sections. Certain areas were anoxic for considerable periods (Hunsrück Schiefer and Wissenbacher Schiefer, for example) and migration of these facies are not documented as in New York.

*Middle Devonian*
The major American Onondaga transgression (D) was progressive rather than sudden and initiation may have preceded the base of the Eifelian as now defined. Similar progressive deepening is indicated in the Ardennes (Fig. 7) but the facies shift is more marked in Devon and Cornwall (House 1975) and it may correspond to a

culmination in the Wissenbacher Schiefer. A major palynological change near the boundary in the Eifel is perhaps a clearer indication of the palaeogeographic change involved. Similar significant changes led to major facies shifts in European Russia (Fig. 7) but precise correlation does not seem to be available. The minor rhythmicity in the Onondaga has an analogy in Belgian equivalents (Lecompte 1970). Later Eifelian change from carbonates to shales is documented in several areas and shows in Cornish basinal argillites (Beese 1982). Perhaps this is a parallel with the Union Springs event (E).

The precision of documentation within the New York Hamilton (F-I) finds no parallel at present in other areas: there are comparisons in the limestone tongues within argillites of the North Devon Devonian.

International documentation of the Tully transgression and Taghanic onlap (Johnson 1970) of North America has been given for Europe and European Russia on an earlier occasion (House 1975) and need not be repeated here. This major event finds widespread recognition around what has been called the Old Red Sandstone Continent and correlated with transgressions leading to carbonate developments as far apart as the Baltic Craton and onto the Precambrian shield areas of Western Australia.

*Upper Devonian*

The rhythmic nature of the New York sequence around the Frasnian (K-O) has a clear analogy in several other areas but precise biostratigraphic work is still needed before synchronicity can be unquestionably claimed. For example, the Western Canada reefs (Klovan 1974) occur in several pulses giving complexes named successively Swan Hills, Leduc and Nisku. Dating of these is given by Johnson (1982) for the Swan Hills as Lowermost *asymmetricus* Zone and drowned in a Middle *asymmetricus* Zone deepening; the Leduc as Uppermost *asymmetricus* Zone or deepening beginning in the Middle *asymmetricus* Zone; and the Nisku, of possible *gigas* Zone age. Difficulties of precise conodont dating in the later New York levels precludes precision but correspondence with late K, L and M/N seem reasonable.

Similar reef pulses are well known in Belgium (Lecompte 1970, Tsien 1975, Burchette 1981). Here, as Mouravieff (1982) has shown, shallow water conodont assemblages do not allow precise correlation with the German conodont zonal standard. Nevertheless the approximate dates are similar to those of Canada, reefs of $F_{2c-d}$ being Lower *asymmetricus Zone;* and $F_{2h}$ reefs, Middle and Upper *asymmetricus* Zone; and $F_{2j}$ Neuville reefs, possible *triangularis* or *gigas* Zone in age.

Some check with New York is given by the entry of a *Shpaeromanticoceras* fauna in the upper Rhinestreet of New York (Kirchgasser and House 1982) since, Sarah Gatlet informs me, similar forms enter just below the basal Matagne of Belgium (which in turn is correlated with the main Kellwasser Kalk levels of Germany).

The German evidence will not be clear until facies shift data in separate sections are analysed separately. (Buggisch 1972) has shown that the main anoxic Kellwasser Kalk horizons span the Lower *gtgas* Zone to *triangularis* Zone interval. These probably correspond to the Rhinestreet and Pipe Creek events (M/N) of New York. But *Crickites holzapfeli* occurs in presumably upper Kellwasser Kalk levels in Germany but in New York such forms are only known in the Hanover Shale.

Transgressions analogous with the Taghanic Onlap are known from European Russia (Fig. 7, House 1975). Descriptions of the Domanik rocks, which are major hydrocarbon source-rock sapropels and bituminous limestones, indicate significant contemporary anoxia. The associated faunas indicate close correlation with the Geneseo (K) of New York. On present, new, definition, the Domanik also straddles the Middle/Upper Devonian boundary.

The basal Famennian Nehden Schiefer suggests a temporary transgressive event and earliest *Cheiloceras* faunas of this level are very widespread internationally. The New York equivalents are the Dunkirk and Gowanda Shales (O/P). This event is immediately followed in many areas by a regressional facies which covers the bulk of the time span of the Cheiloceras Stufe.

During the Upper Famennian, the New York evidence of very extensive clastic facies progradation is matched in Europe with the Pickwell Down and Baggy Beds of Devon, the Condroz sandstones of Belgium, and the Velbert Beds of Germany. In the Baltic States and European Russia facies lapse into non-marine rocks suggesting extensive evaporitic playas; but even these, as Savvaitova (1977) has elegantly shown, carry evidence of finer-scale rhythmicity.

International transgressive pulses so far distinguished in the Famennian are few. The most extensive is that represented by the Annulata Schiefer in Germany (top Platyclymenia Stufe) which is recognisable in eastern North America (event Q), as the Three Forks Shale in north-western U.S.A., and as the strikingly anomalous *Platyclymenia*-bearing cherts at Dugan Pond, California, amid the 1100m thick volcanic complex of the Sierra Buttes Formation (Anderson *et al.* 1974).

In Europe the Wocklumeria Stufe shows deepening facies shifts in the Wocklum Kalk and Hangenberg Schiefer. Precise documentation of these events elsewhere is not so clear; the introduction of Pilton Beds facies in Devon and the marine breaching of the Munster Basin in Ireland (perhaps transgressions T1 and T2 of MacCarthy 1983) may be related to these. Events in the latest Devonian of Ohio (event R) and the Exshaw Shale of eastern North America appear broadly equivalent; both carry quadrate-constricted imitoceratids of the Wocklumeria Stufe and the Exshaw has also *Acutimitoceras-type* nuclei.

## General Comment

Important orogenic events seem to control the broadest features of major Devonian facies shifts and these events include the Caledonian, Acadian and Antler orogenies. The lesser scale rhythmic facies deepenings which have been described here seem inescapably eustatic if the dating of the events internationally is accepted. That there is an international agreement is perhaps surprising. The large-scale control of these may be periods of activity of plate margin constructive activity, and this, as well as modifying sea level would be expected to have sympathetic effect at destructive margins leading to increased subsidence and accretion there; the primary cause of such activity is unknown. The record of the details of facies shift and deepening seems the only way in which this activity can be monitored in time.

As for the finer-scale rhythms, their interpretation is currently elusive. Malenkovitch climatic oscillations, periodic ocean circulation changes resultant upon orbital perturbations, land-based erosion cycles, or smaller-scale eustatic changes are some of the possibilities. Whatever the final interpretation which is accepted their precise documentation in time will, again, depend on refinement in techniques of international biostratigraphic correlation.

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# Planktonic Foraminifera from the Cenomanian of the Wilmington Quarries (S.E. Devon)

M.B. HART

Hart, M.B., 1983. Planktonic Foraminifera from the Cenomanian of the Wilmington Quarries (S.E. Devon). *Proc. Ussher Soc., 5.* 406-410.

The Cretaceous succession of the White Hart Sandpit, Wilmington, S.E. Devon, is described. The distribution of nine species of planktonic Foraminifera is documented and the microfauna correlated with other successions in S.E. England. The most important species recovered are *Rotalipora reicheli* MORNOD and *Favusella washitensis* (CARSEY), the former almost certainly indicating that the Lower/Middle Cenomanian boundary is probably located between the calcareous sands and the overlying sandy limestones.

## Introduction

The village of Wilmington in S.E. Devon lies almost in the centre of a hollow formed by an outcrop of Cenomanian calcareous sands. Exposures of the sands are limited, and today one can only see them in two quarries. The main working quarry to the north west of the village has long been known as the White Hart Sandpit (SY 208999) while to the north east of the village, and now virtually overgrown, lies Hutchins' Pit (ST 216003). Both quarries were mentioned by Jukes-Browne and Hill (1903) largely on account of their rich and varied macrofauna. This initial description has been supplemented by the work of Smith (1961), Smith and Drummond (1962), Kennedy (1970) and Hart (in Durrance and Laming, 1982). This fauna, dominated by echinoids, ammonites, bivalves, and gastropods, can still be collected in the White Hart Sandpit. In *1967* the author visited the quarries to collect micropalaeontological samples, although material from the main face (see Fig. 1) proved to be either too hard, or too deeply decalcified, to yield a significant fauna.

A return visit in 1968 allowed the inspection of a trial pit that was dug into the quarry floor to prove the full thickness of the calcareous sands. The results of these two visits were presented in Hart (1970) and Carter and Hart (1977). More recent visits have provided samples with a better microfauna, and it is this more complete faunal succession that is described here. All the samples used in this work, and their isolated microfaunas, are in the collections of the Department of Environmental Sciences, Plymouth Polytechnic. All the species mentioned in the text, or shown in Figures 2 and 3 are listed in the Appendix.

Figure 1. The main face of Wilmington Quarry (White Hart sandpit) in 1967. The distinctive nodular horizons within the calcareous sands can be seen, as can the sandy limestones above and the overlying white chalk. The calcareous greensand is not present in this face.

## Distribution of Foraminifera

While the foraminiferal fauna of the Wilmington succession is quite rich it is clear that both the specific diversity and the numbers of individuals recovered are profoundly affected by the lithology. Samples were either too hard to process, or were softened by deep, penetrative, decalcification. Gradually however an extensive faunal list has been obtained using some fairly unconventional micropalaeontological processing techniques. Samples have been prepared in the following ways, listed in descending order of usefulness.

> i. grinding under water, in a cast iron pestle and mortar, prior to sieving;
> ii. acid reduction, using dilute HCl;
> iii. inspection of broken rock surfaces under the microscope;
> iv. thin section microscopy.

Because of the variable quality of the products of such methods no statistical treatment of any data is possible and no biometric analysis of the fauna has been attempted. Even so, it must be noted that samples from the calcareous sands contain a higher percentage of planktonic foraminifera than would have been expected at either this stratigraphic level or in such a marginal environment. The zonation developed by Carter and Hart (1977) was based on a combination of planktonic and benthonic species, but with one or two exceptions, none of the diagnostic benthonic species have been found in the succession (see Fig. 2).

However, nine species of planktonic foraminifera have been recorded, and while some of these are of limited stratigraphic value, Others are more significant. *Rotalipora* BROTZEN, 1942 is the most stratigraphically useful genus in the Cenomanian, and in the Wilmington succession two species are represented; *R. appeninica* (RENZ) and *R. reicheli* MORNOD. The latter is very distinctive, being notably piano-convex, with distinct peri-umbilical ridges. Specimens from Wilmington can be favourably compared with material from Sussex, Kent, and Denmark (see Hart 1979). *R. reicheli* has a distinctive stratigraphic range in the U.K. (see Fig. 3) occurring only over a brief interval at, or about, the Lower/Middle Cenomanian boundary, This limited occurrence coincides with a brief re-appearance of the benthonic species *Lingulogavelinella jarzevae* (VASILENKO). The latter species typifies the Lower Cenomanian (Zones 6a-9 of Carter and Hart (1977)) although all over Southern England it re-appears at the Lower/Middle Cenomanian boundary (approximately the Zone 10/Zone 1 li boundary). The co-occurrence of *R. reicheli* and *L.jarzevae,* coupled with the first appearance of *Praeglobotruncana stephani* (GANDOLFI) in the succession, seems to indicate that the calcareous sands belong within Zone 10 - the upper part of the Lower Cenomanian. There is presently no evidence for any part of Lower Cenomanian Zones 7-9. The topmost part of the calcareous sands and the overlying sandy limestones do not contain *R. reicheli,* even though a rich planktonic fauna has been recorded. That fact, together with the occurrence of *Plectina sp. cf P. cenomana* CARTER and HART, 1977, would indicate a position in the lower part of the Middle Cenomanian (Zone 1 Ii).

The other distinctive planktonic species found at Wilmington is *Favusella washitensis* CARSEY. *Favusella* MICHAEL, 1972 is a distinctive genus, very close to *Hedbergella* BRÖNNIMANN and BROWN, 1958 or *Whiteinella* PESSAGNO, 1967, but having chambers covered in a characteristic reticulate ornament. The style of coiling, number of chambers, degree of inflation, and strength of ornament are all highly variable, and because of this any fauna contains a wide range of morphological types. One either accepts this as a function of variability or one produces a plethora of species all from the same stratigraphic level (as done by Michael, 1972). *Favusella* is only found in the U.K. at two levels in the mid-Cretaceous succession. One is at, or near, the Middle/Upper Albian boundary (the same level described by Michael (1972) in Texas) where there is a very wide range of "morphotypes" recorded, while the other is just below the mid-Cenomanian non-sequence (Carter and Hart, 1977). The occurrences in the Wilmington succession correspond with neither of these levels, but the range of "morphotypes" is again quite wide. One is tempted, therefore, to suggest that the development of the reticulate ornament is not so much a function of stratigraphic position as one of ecology. Why the *Hedbergella/Whiteinella* plexus should produce this surface ornament at certain stratigraphic levels, or in certain palaeoenvironments, is not yet understood and work on modern planktonic foraminifera apparently provides no clues.

At the base of the trial pit seen in 1968 were found specimens of *Orbitolina sp. (Orbitolina lenticularis* (BLUMENBACH) of Carter and Hart, 1977). These were found in the pale brown, almost glauconite-free, cross-stratified sandstones shown in Figure 2. They were described in detail by Carter and Hart (1977; pp.17-23, figs 4-6), while Hart, Manley and Weaver (1979) have provided a re-assessment of the age of the fauna. Without further material it is impossible to say more than that the specimens come from close to the Albian/Cenomanian boundary.

The section at Hutchins' Pit is now in an even worse state than it was in 1967/1968, when the material was collected for the earlier investigations. Although better preserved faunas have now been recovered nothing more can be added to the previous descriptions.
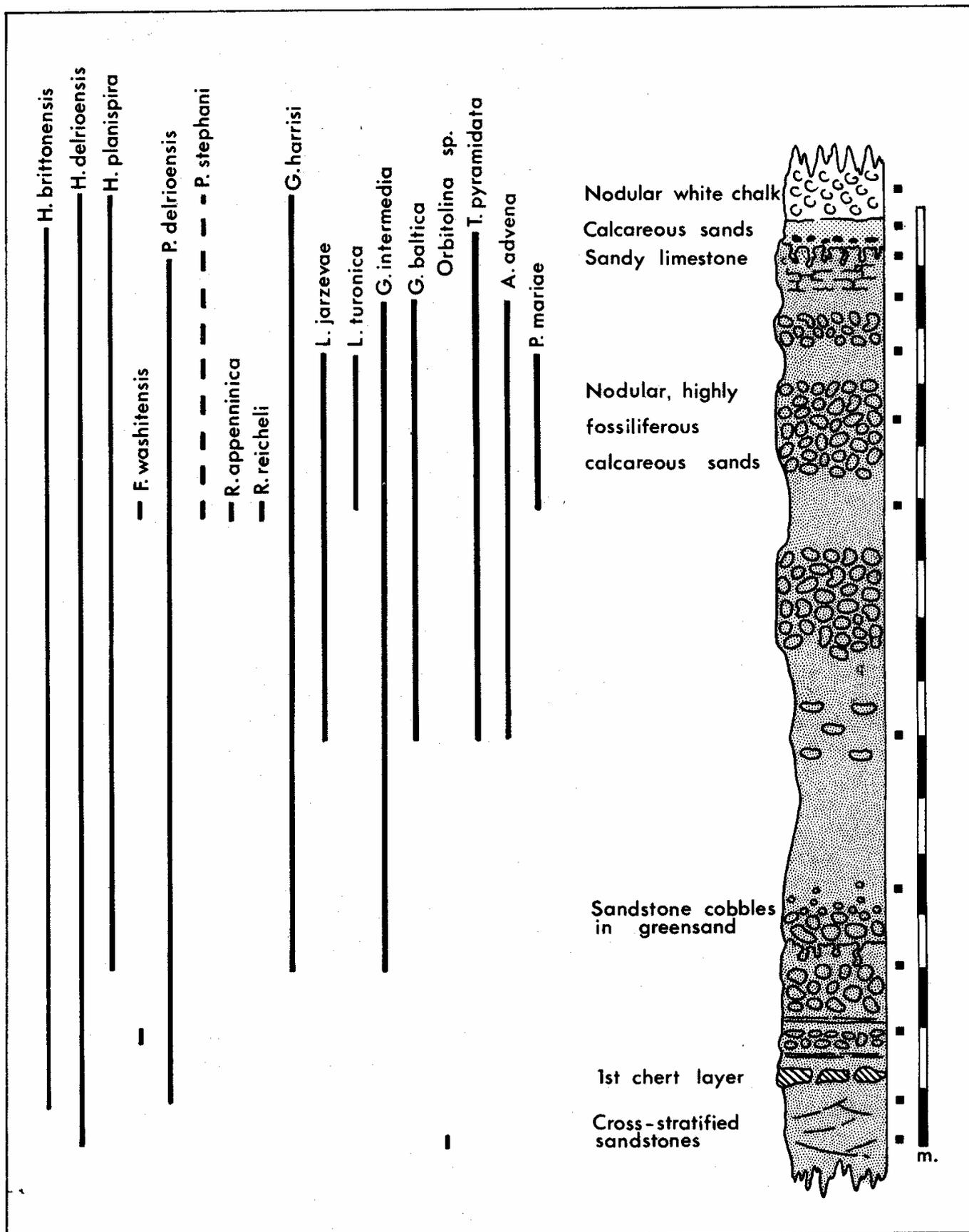
Figure 2. The distribution of selected foraminiferal taxa in the White Hart Sandpit succession.

Figure 3. The correlation of the Wilmington succession with that at Dover. The legend for the Wilmington succession is the same as used in Figure 2.

## Conclusions

On the basis of the faunas described above it is clear that the Cenomanian sands of Wilmington are best placed in the upper part of the Lower Cenomanian (Zone 10), while the overlying sandy limestones are referable to the lower part of the Middle Cenomanian (Zone 1 Ii). This confirms the suggestions made in Carter and Hart (1977), and the correlations with other localities in the area (Hutchins' Pit, Bovey Lane, Beer, and Hooken Cliff) still stand. The calcareous greensand above the prominent erosion surface is equivalent (in part) to Bed C on the coast and in the Bovey Lane Sandpit. None of these conclusions contradict the published macrofaunal data (Kennedy, 1970; Rawson et al. 1978), although there is currently no microfaunal evidence for extending the sands succession down to the base of the Cenomanian (as done by Rawson et al.).

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# Appendix

*Favusella washitensis* (CARSEY) = *Globigerina washitensis* CARSEY 1926, p 44, pi 7, fig 10, pi 8, fig 2.

*Guembilitria harrisi* TAPPAN, 1940, p 115, pl 19, fig 2a-b

*Hedbergella brittonensis LOEBLICH and TAPPAN. 1961, pp 274-275 pl 4, figs 1-8*

*Hedbergella delrioensis* (CARSEY) = *Globigerina cretacea* d'Orbigny var. *delrioensis* CARSEY 1926, p 43.

*Hedbergella planispira* (TAPPAN) = *Globigerina planispira* TAPPAN, 1940, p 12, pl 19, fig 12.

*Praeglobotruncana delrioensis* (PLUMMER) = *Globorotalia delrioensis* PLUMMER, 1931, p 199, pi 13, fig 2a-c.

*Praegtobotruncana stephani* (GANDOLFI) = *Globotruncana stephani* GANDOLFI, 1942, p 130, pl 3, figs 4,5, pi 4, figs 36, 37,

41-45, pi 6, figs 4,6, pi 9 figs 5,8, pi 13, fig 5, pi 14, fig 2.

*Rotalipora appennintca* (RENZ) = *Globotruncana appeninica* RENZ, 936, p 14, fig 2.

*Rotalipora cushmani* (MORROW ) = *Globorotalia cushmani* MORROW 1934, p 199, pi 31, figs 2,4.

*Rotalipora greenhornensis* (MORROW) = *Globorotalia greenhornensis* MORROW, 1934, p 199, pl 39, fig 1.

*Rotalipora reicheli* (MORNOD) = *Globotruncana (Rotalipora) reicheli* MORNOD, 1950, p 583, fig 5 (Wa-c).

*Gavelinella baltica* BROTZEN, 1942, p 50, pl 1, fig 7.

*Gavelinella interraedia* (BERTHELIN) = *Anomalina intermedia* BERTHELIN, 1880, p 67, pi 4, fig 14a-b.

*Lingulogavelinella jarzevae* (VASILENKO) = *Cibicides (Cibicides) jarzevae* VASILENKO, 1954, p 121, pi 17, fig 3a-c.

*Arenobulimina advena* (CUSHMAN) = *Hagenowella advena* CUSHMAN, 1936, p 43, pl 6, fig 21a-b.

*Arenobulimina advene* (CUSHMAN) = *Hagenowella advena* CUSHMAN, 1936, p 43, pi 6, fig 21a-b.

*Plectina cenomana* CARTER and HART, 1977, pp 12-13, pl 2, fig 9.

*Plectina mariae* (FRANKE) = *Gaudryina ruthenica* Reuss var. *mariae* FRANKE, 1928, p 146, pi 13, fig 15a-b. *Pseudotextulariella cretosa* (CUSHMAN) =*Textulariella cretosa* CUSHMAN. 1932, p 97, pi 11, figs 17-19.

*Tritaxia pyramidata* REUSS, 1862. p 32, 88, pl 1, fig 8a-c.

# Disseminated tin sulphides in the St Austell granite

D.A.C. MANNING*

Manning, D.A.C., 1983. Disseminated tin sulphides in the St Austell granite. *Proc. Ussher Soc., 5,* 411-416.

The tin sulphides stannite ($Cu_2(Fe,Zn)$ $SnS_4$), stannoidite ($Cu_8(Fe,Zn)_3Sn_2S_{12}$) and mawsonite ($Cu_6Fe_2SnS_8$) occur with cassiterite as a dissemination within the Hensbarrow stock of non-megacrystic lithium-mica granite in the composite St Austell granite body. Two parageneses are present: one consisting of the assemblage stannite - stannoidite, and the other of stannoidite -bornite-mawsonite. Available experimental data suggest that these crystallised below 500°C, with mawsonite forming below 390°C by reaction between stannoidite and bornite. A possible crystallisation sequence with increasing sulphur fugacity and decreasing temperature is proposed: cassiterite → stannite + stannoidite:- →  stannoidite + bornite. → mawsonite. The tin sulphide mineralisation is apparently unrelated to occurrences of wolframite, cassiterite and columbite within pegamatitic vugs at the same locality, and also predates jointing. Supergene alteration resulted in the formation of varlamoffite and secondary copper phosphates (turquoise and rashleighite). It is postulated that the larger Nanpean stock of non-megacrystic lithium-mica granite may also have been similarly mineralised, and if so may have acted as a source area for a proportion of the superficial tin deposits which lie to the north of the St Austell granite.

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## Introduction

This paper describes disseminated tin sulphide mineralisation which occurs within the Hensbarrow stock of non-megacrystic lithium-mica granite in the composite St Austell granite body (Fig. 1). The tin sulphide occurrence is currently well exposed in active china-clay workings; full details of the locality can be obtained from the Production Department of English Clays, Lovering, Pochin and Co., Ltd (John Keay House, St Austell, PL25 4DJ), to whom all enquiries should be addressed.

Several tin sulphide mineral species are known, and early confusion over names and chemical compositions has largely been eliminated (Petruk 1973; Moh 1975; Kissin and Owens 1979). Stannite (($Cu_2(Fe,Zn)$ $SnS_4$) has been recorded from a number of localities in Cornwall, including the Hensbarrow area (Russell and Vincent, 1952). The other tin sulphides mawsonite ($Cu_6Fe_2SnS_8$) and stannoidite (alternatively named hexastannite; ($Cu_8(Fe,Zn)_3Sn_2S_{12}$) are known to occur in south-west England, but very few detailed descriptions are available. By describing the occurrence of these minerals in the Hensbarrow stock, it is hoped that further discoveries may be made so that future studies may help both to clarify some of the relationships shown by the tin sulphide minerals overall as well as to place some constraints on the factors controlling their crystallisation.

## The field relations of the tin sulphide occurrence

Stannite, mawsonite and stannoidite occur disseminated within non-megacrystic lithium-mica granite which makes up the small Hensbarrow stock, outcropping about 2.5km SSE of Roche. This stock and its intrusive contact are well exposed in a number of working and redundant china-clay pits, and records of underground exposures in now inaccessible mine workings are also available (Ussher et al. 1909; Russell 1948; Russell and Vincent 1952; Dines 1956). Within the St Austell granite, the non-megacrystic lithium-mica granite facies carries quartz, orthoclase, albite (An0-3), zinnwaldite or lepidolite, and topaz as essential silicate minerals; ilmenorufile (Fe, Nb-rich rutile) occurs as tiny inclusions (up to 5 x 20  ) within the lithium mica. Pegmatitic rugs (up to 50cm) and sub-horizontal layers (up to 30cm) are particularly common in the Hensbarrow stock, but are relatively rare, apart from near contacts, in the larger body which outcrops near Nanpean (Fig. 1). By analogy with the Tregonning granite (Stone 1975) the pegmatite complex which characterises the Hensbarrow stock is taken as evidence that the present exposures are of the roof zone of this body. The pegmatitic vugs carry a very varied mineralogy. In addition to quartz, tourmaline and muscovite (gilbertite), apatite, microcline, orthoclase and zinnwaldite are common; Hodkinson and Clark (1977)
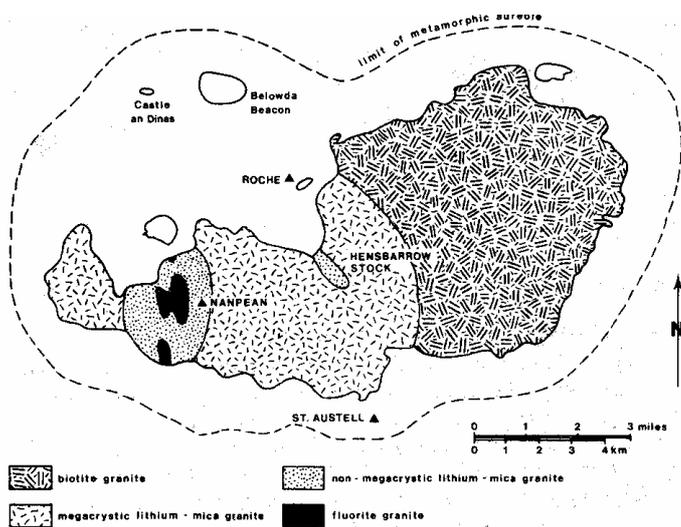
Figure 1. Outline geological map of the St Austell granite, based on the updated map by Manning and Exley (in press).

report columbite, and Barstow (1982) records wolframite, cassiterite, stannite, arsenopyrite, varlamoffite, opal, torbernite and variscite. The pegmatitic vugs also locally contain bornite, chalcopyrite and rashleighite (a secondary copper phosphate), which have been observed in the course of this study. The pegmatitic minerals have not been observed to occur together within a single bug-typical parageneses include:

    quartz - tourmaline - muscovite (the most common)
    quartz - wolframite - zinnwaldite
    quartz - bornite - chalcopyrite
    quartz - cassiterite – rashleighite.
Similarly, joints carry a number of different minerals; fluorite is common, and both turquoise and varlamoffite form a joint fill.

The tin sulphide minerals occur as spots up to 2cm across of dark, metallic lustre, grains mixed with silicates. Individual grams range in size up to 1mm across. The tin sulphides are restricted to working faces in several of the upper levels of an active china-clay pit; they form approximately up to 1% of the rock along a total face length of several hundred metres, with a vertical extent in excess of 50m. Joints carrying varlamoffite and turquoise are particularly common within the tin sulphide - bearing granite.

In the Hensbarrow stock, tin sulphide mineralisation has previously been exposed in mine workings which probably intersected the north-eastern contact between non-megacrystic lithium mica granite and megacrystic lithium-mica granite. Several closely spaced thin cassiterite-stannite-wolframite bearing veins, trending 040°E, were at one time worked, both in open pits as well as underground. According to Ussher *et al.* (1909) and Russell (1948) the veins ranged up to 15cm across (exceptionally up to 50cm), and were usually not more

than 30cm apart. The veins, which were greisen-bordered and separated by heavily kaolinised granite, contained quartz, kaolin, cassiterite, wolframite, rashleighite? tourmaline and gilbertite mica as well as stannite. Fluorite is recorded as a joint coating in the adjacent granite, and topaz is reported to occur within the greisen. Dines (1956) adds that the country rock adjacent to the vein was often observed to be impregnated with cassiterite. In addition, Russell and Vincent (1952) describe varlamoffite from a vein at outcrop and dump material, closely associated with the copper phosphates rashleighite and turquoise as well as cassiterite, wolframite and stannite. These particular mine workings are well known as a stannite occurrence (Russell and Vincent, 1952) and so it was considered that the stannite group minerals currently exposed in the working china-clay pit may be related to the lode system of the old mine; however, detailed mapping has shown that the present exposures do not lie along the strike of known lodes (which in any case appear to die out before reaching the china-clay pit) unless these are offset by faulting.

## The sulphide mineral assemblage

The tin sulphide species were identified by a combination of electron microprobe and petrographic techniques. In addition to stannite, stannoidite and mawsonite, bornite ($Cu_5FeS_4$), digenite ($Cu_{1.8}S$), tennantite ($Cu_{10}(Fe,Zn)_2As_4S_{13}$) and cassiterite ($SnO_2$) have also been identified. Quantitative analyses of stannite, stannoidite and mawsonite are given in Table 1. The analytical conditions and standards used are as follows: 1) at Manchester, for stannite: 15 kV accelerating voltage, $0.3 \times 10^{-8}$ A beam current, point analysis; 2) at Nancy, for stannoidite and mawsonite: 20 kV accelerating voltage, $0.7 \times 10^{-8}$ A beam current, 5 p raster area of analysis. In both cases the following standards were used: for Zn, As, Sn - pure metals; for Fe and S - pyrite; for Cu - chalcopyrite. Despite the use of these standards (ideally: synthetic Cu - Fe - Zn - Sn sulphides should be used - Kissin and Owens 1979) reasonably good stoichiometry was obtained, and then analyses agree closely with those given by other authors (Springer 1968; Petruk 1973).

## Textural relationships

The optical properties of the tin sulphide minerals described here are summarised and compared by Oen (1970), on whose work the following summary is based, for consistency in the description of colours. Stannite, stannoidite and mawsonite all have reflectivities qualitatively similar to bornite. Stannite shows moderate bireflectance, is light grey in colour (in air) and shows moderate anisotropy. Stannoidite shows strong bireflectance, is greyish brown in air (darker than stannite) and is strongly anisotropic (with violet blue to orange red polarisation colours); it also clearly shows twinning. Mawsonite is strongly bireflectant, is greyish orange brown in air (darker than stannoidite) and is strongly anisotropic (orange yellow to blue polarisation colours).

|        | stannite |       | stannoidite |        | mawsonite |       |
|--------|----------|-------|-------------|--------|-----------|-------|
|        | a        | b     | a           | b      | a         | b     |
| Cu     | 31.37    | 2.163 | 39.42       | 8.067  | 44.07     | 6.027 |
| Fe     | 8.31     | 0.652 | 9.79        | 2.278  | 12.85     | 1.999 |
| Zn     | 5.14     | 0.341 | 3.27        | 0.651  |           | -     |
| Sn     | 25.99    | 0.953 | 18.19       | 1.993  | 13.35     | 0.978 |
| S      | 28.47    | 3.892 | 29.62       | 12.011 | 29.50     | 7.996 |
| totals | 99.32    | 8.000 | 100.28      | 25.000 | 99.77     | 17.000 |
|        |          |       |             |        |           |       |
| Fe + Zn |         | 0.993 |             | 2.929  |           | 1.999 |
| Fe/(Fe+Zn) |      | 0.657 |             | 0.778  |           | 1     |
| Sn:S   |          | 1:4.084 |           | 1:6.027 |          | 1:8.176 |

Table 1. Compositions of tin sulphides from the Hensbarrow stock, St Austell granite, Cornwall. Stannite determinations were made using the Cameca Camebax/Link Systems energy dispersive system at Manchester University; stannoidite and mawsonite determinations were made using the automated Cameca Camebax wavelength dispersive system at the University of Nancy I. Column a gives the weight percentage for each element, and column b the atomic proportions, calculated to the appropriate number of atoms, in each case.
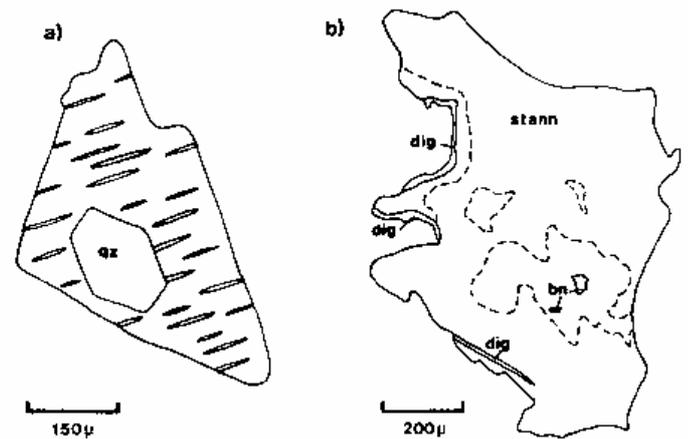


Figure 2. Typical relationships shown by the tin sulphides from the Hensbarrow stock. (a): stannite, with exsolution lamellae of stannoidite. Note the euhedral section presented by quartz. (b): stannoidite (stann) and bornite (bn) with mawsonite (enclosed by broken lines) and digenite (dig).

The tin sulphides rarely occur as isolated grains. Two parageneses have been recognised. In the first, stannite and stannoidite are associated - they share planar grain boundaries within irregular composite grains (up to lmm) whose overall shape is determined by the surrounding silicate minerals. Stannoidite additionally occurs as exsolution lamellae within stannite (Fig. 2a). Some grains of stannoidite carry euhedral inclusions of muscovite; these may have crystallised earlier and subsequently become included during stannoidite crystallisation, or may have crystallised at the same time as the stannoidite. The second paragenesis is mineralogically more complex. Mawsonite characteristically occurs as irregular rims between bornite and stannoidite, as well as an irregular replacement of stannoidite, extending inwards from the margins of composite grains (Fig. 2b). Both mawsonite and stannoidite are cut by veins of digenite, which also forms a rim together with tennantite in certain cases. In particular, in the second paragenesis, grains of tennantite replacing digenite occur, and these are in turn replaced by muscovite.

In addition to the tin sulphides, cassiterite occurs in association with the stannite - stannoidite assemblage as euhedral inclusions within stannite and as irregular isolated grains up to 750$\mu$ across.

The sulphide assemblages appear m certain examples to have crystallised within vugs; for example, the shape of individual sulphide grains is frequently seen to be controlled by euhedral quartz crystals (Fig. 2a). In some cases, the primary lithium mica is clearly seen to be replaced by the sulphides.

## Discussion

The system Cu-Zn-Fe-Sn-S is extremely complex, and to date very few experimental studies have been carried out to determine phase relationships for four and five component compositions. However, two studies in particular provide constraints which allow the observed mineralogical and textural relationships to be interpreted. Lee *et al.* (1975) have examined phase relationships in the Zn-free system, Cu-Fe-Sn-S, to determine the conditions required for the synthesis of stannoidite and mawsonite. Within this system, Lee *et al.* (1975) report a number of invariant assemblages, including at 500°C:

   stannoidite - stannite - chalcopyrite - bornite

and at 390°C, at which temperature mawsonite is stable:

   bornite - digenite - stannoidite - mawsonite.

Mawsonite was found to decompose on heating above 390°C to give stannoidite, bornite and, pyrite or chalcopyrite. Lee *et al.* (1975) report that zinc-free stannoidite is stable up to 830°C, at which temperature it decomposes to give bornite, stannite and chalcopyrite, However, the presence of approximately 3 wt% Zn in stannoidite reduces the upper stability limit to around 500°C (Lee *et al., op cit).* On the other hand, Springer (1972) has shown that stannite-kesterite ($Cu_2FeSnS_4$ - $Cu_2ZnSnS_4$) solid solutions are stable up to 878°C ($Cu_2FeSns_4$) - 1002°C ($Cu_2ZnSns_4$). In addition, these compositions have a miscibility gap below 680°C (providing a potential geothermometer) arising from a structural inversion which occurs at reduced temperatures with increasing zinc content (Springer, 1972). According to early experimental work, the high temperature form is cubic, and the low temperature form tetragonal (Bernhardt, 1972; Kissin and Owens, 1979)
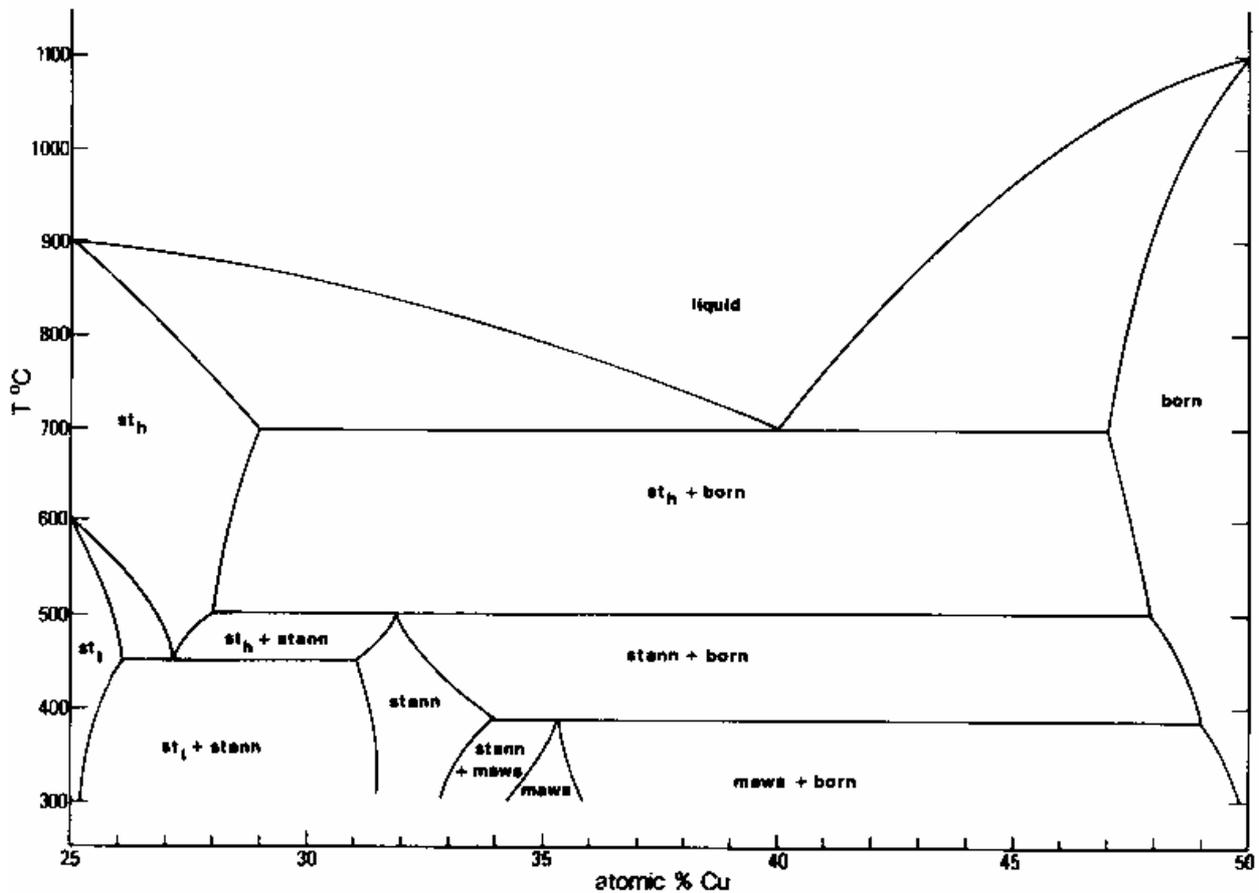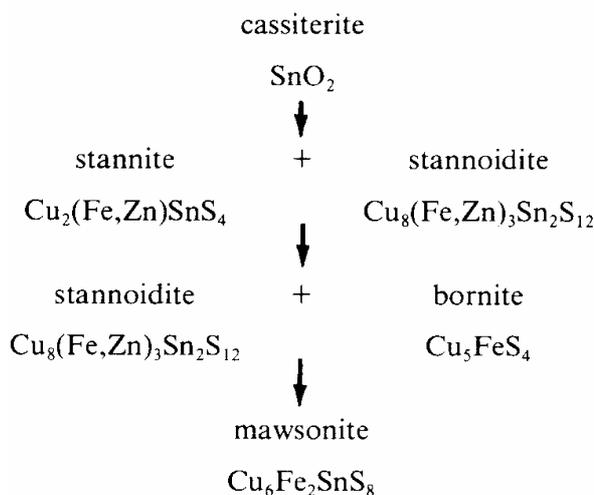
Figure 3. Hypothetical phase relationships in the pseudobinary system stannite-bornite, for zinc-bearing stannite and stannoidite. Symbols: sth = high stannite polymorph; st, = low stannite polymorph; stann = stannuidite; maws = mawsonite; born = bornite.

but more recent studies for the zinc-free composition $Cu_2FeSnS_4$ have shown that the high form is tetragonal and the low form cubic (Wang, 1982). Further work remains to be done to clarify the details of the inversion (especially for zinc-bearing compositions) which only involves slight structural changes. Experimental data are not available for the stannite - stannoidite exsolution relationships, although these have been described in the literature (e.g. Picot and Johan, 1977).

The phase equilibria which can be used to discuss the origin of the tin sulphide parageneses observed in the Hensbarrow occurrence are summarised schematically in Figure 3. This is based on the pseudobinary section stannite-bornite presented by Oen (1970) with some modifications derived from more recent experimental studies (Springer, 1972; Lee *et. al.* 1975; Kissin and Owens, 1979; Wang, 1982) so that zinc-bearing compositions can be considered. It must be stressed these phase relationships were essentially derived by Oen from the observation of natural mineral assemblages coupled with constraints provided by existing experimental studies. Reference temperatures derived from experimental studies include: 1) the stannite inversion at

about 600°C; 2) the breakdown of zinc-bearing stannoidite at approximately 500°C; and 3) the breakdown of mawsonite at 390°C. The compositional limits for solid solution between phases have not been determined. Using this diagram, possible crystallisation histories for the two observed phase assemblages can be deduced. First of all, the stannite - zincian stannoidite assemblage, which contains no bornite, must have crystallised (or completely recrystallised) below about 500°C. The observation that both species occur as grains of similar size, often sharing planar grain boundaries, suggests that the two may have crystallised together. The second assemblage is of zincian stannoidite and bornite which are replaced by mawsonite; again, these crystallised below 500°C, with the formation of mawsonite taking place below about 390°C. The difference between the two assemblages essentially lies in their crystallisation on either side of the stannoidite composition (Fig. 3), This may be controlled by the sulphur fugacity which obtained during crystallisation, as Lee *et al.* (1975) observed that the formation of mawsonite is favoured by an increase in sulphur fugacity (in comparison with that in equilibrium with stannoidite - bornite assemblages) as well as by a reduction in temperature. It is therefore possible to

postulate a crystallisation sequence for the tin minerals, in order of increasing sulphur fugacity and/or decreasing temperature:

$$\text{cassiterite}$$
$$\text{SnO}_2$$
$$\downarrow$$

| stannite | + | stannoidite |
|---|---|---|
| $\text{Cu}_2(\text{Fe},\text{Zn})\text{SnS}_4$ | | $\text{Cu}_8(\text{Fe},\text{Zn})_3\text{Sn}_2\text{S}_{12}$ |

$$\downarrow$$

| stannoidite | + | bornite |
|---|---|---|
| $\text{Cu}_8(\text{Fe},\text{Zn})_3\text{Sn}_2\text{S}_{12}$ | | $\text{Cu}_5\text{FeS}_4$ |

$$\downarrow$$
$$\text{mawsonite}$$
$$\text{Cu}_6\text{Fe}_2\text{SnS}_8$$

This is consistent with the observed parageneses, and would account for the occurrence of the two observed sulphide phase assemblages.

The occurrence of varlamoffite and secondary copper phosphates (rashleighite and turquoise; Russell, 1948) as a joint fill in the tin sulphide-bearing granite probably arises from the supergene oxidation of the tin sulphides, as this assemblage is a typical decomposition product of stannite (Sharko, 1971).

From a wider point of view, the formation of the disseminated tin mineralisation may be related to large scale post-magmatic processes which are believed to have affected the St Austell granite. An initial, pre-joint, stage of fluid circulation, involving essentially meteoric water, is believed to have initiated the release of fluorine from primary F-bearing silicates (topaz and lithium mica) to the hydrothermal fluid (Manning and Exley, in press). This type of alteration occurs within the Hensbarrow stock of non-megacrystic lithium-mica granite, including the area which is host to the tin-sulphide dissemination. Fluid circulation is considered to have involved principally grain boundaries prior to joint formation (Durrance *et al.* 1982; Manning and Exley, in press). The fluid pathways frequently changed, because of recrystallisation of quartz in particular, and so fluid access to a given volume of rock also changed with time. This may account for the very varied mineralogy of in particular the pegmatitic vugs described previously, some of which may have had access to fluids of a given composition while others were isolated. As well as fluid-rock interaction within the granite, reaction with the country rock will also have affected the fluid composition - this in particular may have affected the sulphur fugacity.

It is quite possible that the Nanpean stock of non-megacrystic lithium-mica granite also had a roof complex similar to that currently exposed in the outcrop of the Hensbarrow stock, but that this has now been removed by erosion. By analogy with the Hensbarrow stock,

pegmatitic and disseminated cassiterite, wolframite and columbite may have occurred within the Nanpean stock roof zone, and may have contributed, in addition to cassiterite and wolframite from veins, to the superficial cassiterite deposits which lie to the north of the St Austell granite body.

## Conclusions

The observed phase assemblages and available experimental data suggest that the tin minerals crystallised in a sequence governed by either increasing sulphur fugacity or decreasing temperature or a combination of the two. Crystallisation of stannite and stannoidite within vugs followed cassiterite, and may have taken place initially in the temperature interval 500°C - 390°C, together with sericitic alteration, but continued below this with the formation of mawsonite. Fluid pathways appear to have involved grain boundaries (and possibly interconnecting vugs) to a greater extent than in nearby stannite - cassiterite occurrences - this appears to have been a local characteristic of the host granite within this part of the roof zone, as most mineralisation in this area occupies planar fractures. Joint formation within the granite host appears to postdate the disseminated tin sulphide mineralisation, but ultimately permitted the supergene formation of varlamoffite and secondary copper phosphate minerals.

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# An experimental study of the effects of Lithium on the granite system

JOANNA S. MARTIN

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Preliminary results are given for the system Albite (Ab)-Orthoclase (Or)-Quartz (Qtz) with the addition of 1 and 2 wt % $Li_2O$ at 1 kb with excess water. The effects of small amounts of Li on the granite system are compared with those of fluorine showing that although both result in the lowering of liquidus temperature, Li does not produce the shift in field boundary and enlargement of the quartz field shown by fluorine. $Li_2O$, F and $P_2O_5$ values are given for some Cornish granites and a comparison is made between the composition of some Li bearing and Li free micas in the Cornish granites.

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## Introduction

In their study of the synthetic granite system Ab-Or-Qtz, Tuttle and Bowen (1958) pointed out that granites are eminently suitable for experimental study on account of their simple major oxide composition. They determined the position of the field boundary (Fig. 1) and the minimum which had a composition of 34% Ab, 29% Or and 37% Qtz and was at a temperature of 730°C at 1 kb $PH_2O$. Interest then moved to the addition of other volatiles to either natural or synthetic granites. Wyllie and Tuttle (1959, 1961, 1964) investigated the effects of $CO_2$, $NH_3$, HF, $SO_3$, $P_2O_5$, HC1 and $Li_2O$ on granites. This work was followed by Chorlton and Martin (1978) who studied the effect of B on the granite solidus. Manning (1982) added F to synthetic granites to determine its effect on liquidus temperatures (Fig. 1). This work showed that at 1 kb $PH_2O$ the addition of 1 wt % F resulted in a 40°C decrease in minimum temperature, enlargement of the quartz field and the movement of the minimum towards more Na-rich compositions. With the addition of 4 wt % F there was a further 60°C drop in minimum temperature, a greater enlargement of the quartz field and also further movement of minimum composition towards the albite apex. Manning suggested that the lowering of the liquidus temperature with the addition of F was the result of depolymerization of the melt and that the shift in field boundary was caused by preferred complexing of Al, thus explaining the enlargement of the quartz field.

The Cornish granites, and in particular the St Austell granite, are known to have fairly high bulk $Li_2O$ contents, some containing over 1 wt % $Li_2O$ (Table 1). Experimental work on the addition of $Li_2O$ to natural granites (Wyllie and Tuttle, 1964) showed a marked lowering in solidus temperature with the addition of small amounts of $Li_2O$. The effects of Li on the liquidus temperature and the phase relations in the system Ab-Or-Qtz seemed worthy of study and are the topic of this paper.
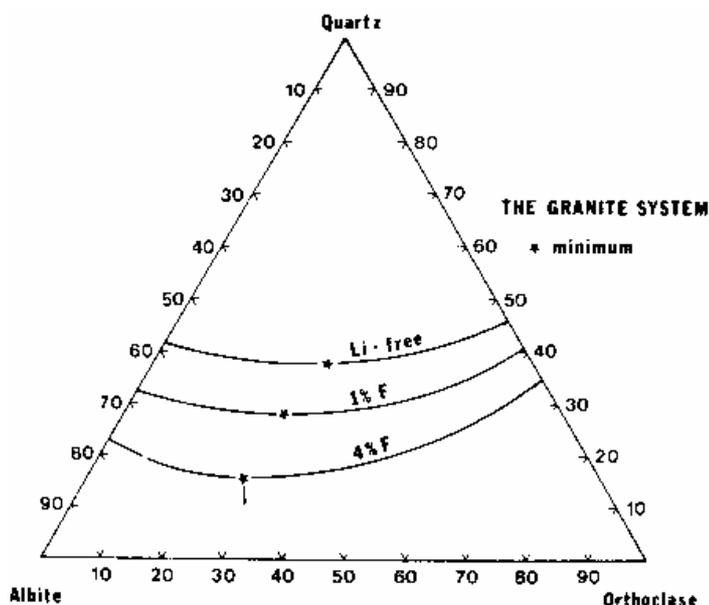


Figure 1. The granite system, with field boundaries for the Li and F free system and the system with the addition of 1 and 4 wt % F, with excess H20 at 1 kb (showing alkali feldspar solvus at 4 wt % F).

## Experimental Methods

Synthetic gel starting materials were made up by the method of Hamilton and Henderson (1968). After the initial firing of the gel the Li was added as a nitrate solution and the mixture fired again and thoroughly ground and homogenized. The lithium content and homogeneity of the gel were checked chemically. A number of starting compositions were made in order to determine the position of the field boundary. The first set of experiments used gels to determine the liquidus temperatures, which were then checked in reversal experiments using fused and crystallized gels as starting material

In each experiment about 5 wt % $H_2O$ was added to the starting material and sealed in a platinum capsule before running for one week at 1 kb in a Tuttle cold seal pressure vessel. The temperature was monitored regularly with an external thermocouple. At the end of one week the run was quenched and the contents of the capsules examined optically and by X-ray diffraction techniques.

## Results

Initial experiments were conducted with the addition of 1, 2 and 4 wt % $Li_2O$, but with the latter it was found that the lithium aluminosilicate petalite appeared on the liquidus, so these compositions were abandoned as lying outside the granite system. The only other phases observed in the charges were quartz, occurring as bipyramidally - shaped crystals, and laths of alkali feldspar. The petalite, when present, formed mats of needle-like crystals. The glasses were always clear and frequently contained vesicles which were sometimes lined with crystals formed from the vapour phase during quenching.
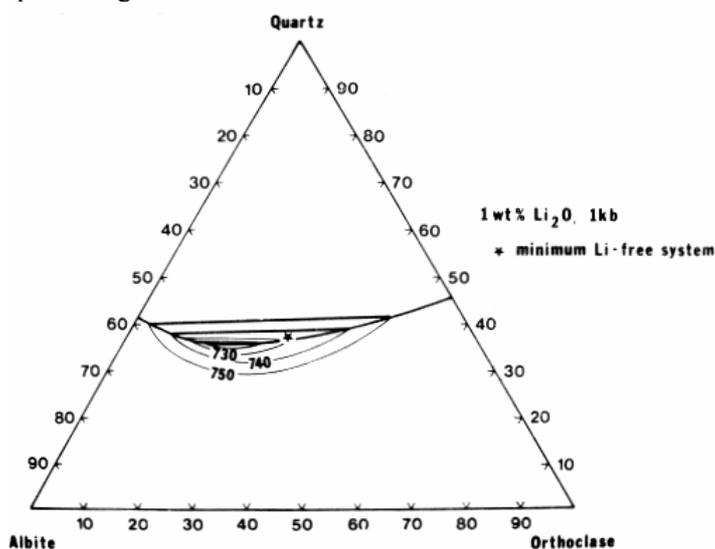


Figure 2. The granite system with the addition of 1 wt % $Li_2O$ with excess $H_2O$ at 1 kb.

The phase diagram for the addition of 1 wt % $Li_2O$ (Fig. 2) suggests that there is little movement of the field boundary relative to the Li-free system but that there is a marked drop in the temperature of the minimum which the results indicate to be less than 700°C, at least 30° below that of the Li-free system. The phase diagram for the addition of 2 wt % $Li_2O$ (Fig. 3) also shows little significant movement of the field boundary but there is evidence for a further drop in minimum temperature. The geometry of the phase diagram suggests that the minimum composition is displaced to rather more Ab rich composition than in the Li-free system although the actual composition of the minimum has not yet been precisely determined.
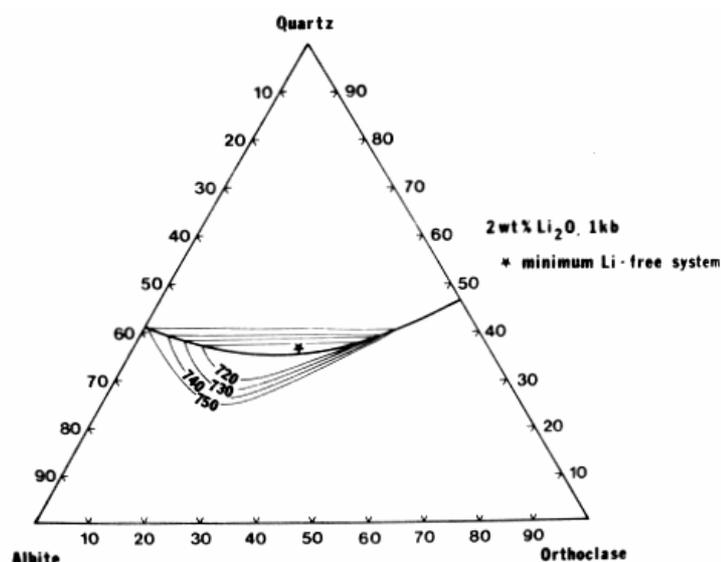


Figure 3. The granite system with the addition of 2 wt % $Li_2O$ with excess $H_2O$ at 1 kb.

## Petrological Work

The eastern part of the St Austell granite consists mainly of the megacrystic biotite granite. The western part is made up of the megacrystic and non-megacrystic lithium mica granites with some small areas of fluorite granite (Exley and Stone, 1964; 1982). The other two important lithium rich outcrops are the Tregonning-Godolphin granite and the Meldon aplite, north of Dartmoor.

Bulk rock compositions of the rocks from the St Austell granite have been determined and some selected results are given in Table 1. $P_2O_5$ values were obtained by X-ray fluorescence, $Li_2O$ values by atomic absorption and the F values are previously published.

The Meldon aplite has very high $Li_2O$ values as well as high F and moderately high $P_2O_5$. The Gunheath stock of the non-megacrystic Li-mica granite has an average $Li_2O$ value of over 1% coupled with very high F and $P_2O_5$. In the St Austell granite, the main outcrop of the non-megacrystic lithium mica granite has lower values of all three components and on undergoing hydrothermal alteration to form the fluorite granites there is a very sharp drop in the amount of $Li_2O$ present. The megacrystic Li-mica granite has lower values of all three components and the values for the biotite granite are included for comparison.

In some rocks there is not sufficient Ca-present to combine with all the $P_2O_5$ to form apatite and it seems likely that one or more other phosphates are present, the most likely being amblygonite $LiAl(F,OH)PO_4$.

|  | $Li_2O$ Wt % | F wt % | $P_2O_5$ Wt % |
|---|---|---|---|
| Meldon aplite | 0.62 | 1.39 | 0.49 |
|  | 0.43 - 1.88 (3) | Edmonds et al. (1968) | 0.47 - 0.53 (2) |
| Non-megacrystic Li-mica granite i) Gunheath |  |  |  |
|  | 0.59 | 1.78 | 0.9 |
|  | 0.55-0.63 (5) | Manning (1979) | 0.67-1.25 (6) |
| ii) Main outcrop Kernick, Trevisco, Slip, Goonvean, Prosper | 0.36 | 1.38 | 0.56 |
|  | 0.11-0.39 (5) | Manning (1979) | 0.50-0.59 (5) |
| iii) Rocks containing fluorite | 0.01 (4) | 1.23 | 0.81 |
|  |  | Manning (1979) | 0.70-1.11 (6) |
| Megacrystic Li-mica granite | 0.22 | 0.38 | 0.58 |
|  | 0.09-0.50 (12) | Exley (1959) | 0.37-0.91 (11) |
| Biotite granite (Luxullian) | 0.07 | 0.24 | 0.35 (1) |
|  | Exley & Stone (1982) | Exley & Stone (1982) |  |

Table 1 Bulk Rock Analyses - St Austell Granite and the Meldon Aplite (Figures in brackets refer to the number of analyses).

In most rocks the Li is concentrated in the micas and preliminary results from over thirty samples of mica from the St Austell granite show that the majority of the primary micas are zinnwaldites. This species of Li-mica is characterized by 5-15 wt % FeO approximately 20 wt % $Al_2O_3$ and around 4 wt % $Li_2O$, the F content of these micas is usually > 5 wt %.

Table 2 gives the analyses of micas from three localities, all elements were obtained on the electron microprobe, except Li20 values which were obtained by Dr R. Mason using the ion microprobe. Micas 5000 and 5026 fall within the zinnwaldite compositional range defined above. It is interesting to note that zinnwaldites are trioctahedral micas which frequently have deficiencies in the octahedral site, as in sample 5000 (Table 2). However zinnwaldite mica 5026 has more than six ions in the octahedral sites which may be the result of analytical error and hence the erroneous calculation of the cell formula. The presence of significant amounts of other elements such as Rb would tend to increase the anion sum and so decrease the number of cations. The mica 5017 is a recrystallized muscovite from a partially kaolinized Li mica granite, the $Li_2O$ content has not been determined but is unlikely to be significant. The low total and low F content together indicate the presence of substantial amounts of $H_2O$ probably about 4 wt %. This mica appears to be a stoichometric muscovite. Chaudhry and Howie's work (1973) on the Meldon aplite shows the micas from that locality to be lithian muscovites or lepidolites with $Li_2O$ values of 3.6-6.4 wt % and very low FeO content.

|  | TREVISCO 5000 | | GUNHEATH 5026 | | PROSPER 5017 | |
|---|---|---|---|---|---|---|
| $SiO_2$ | 45.97 | | 48.60 | | 47.20 | |
| $Al_2O_3$ | 20.63 | | 20.26 | | 35.55 | |
| $TiO_2$ | 0.06 | | 0.17 | | 0.28 | |
| FeO | 11.78 | | 7.87 | | 0.40 | |
| $Li_2O$ | 3.97 | | 5.11 | | - | |
| MnO | 0.06 | | 0.64 | | - | |
| MgO | 0.11 | | - | | 0.62 | |
| CaO | - | | - | | - | |
| $Na_2O$ | 0.09 | | 0.34 | | 0.67 | |
| $K_2O$ | 9.84 | | 10.01 | | 10.40 | |
| F | 6.39 | | 6.97 | | 0.33 | |
|  | 98.90 | | 99.97 | | 95.45 | |
| O≡F | 2.69 | | 2.93 | | 0.14 | |
| Total | 96.21 | | 97.04 | | 95.31 | |
| Si | 6.623 | 8.000 | 6.793 | 8.000 | 6.248 | 8.000 |
| Al | 1.377 | | 1.207 | | 1.752 | |
| Al | 2.121 | | 2.135 | | 3.794 | |
| Ti | | | 0.017 | | 0.028 | |
| Fez+ | 1.420 | 5.871 | 0.924 | 6.024 | 0.045 | 3.988 |
| Li | 2.304 | | 2.872 | | | |
| Mn | | | 0.076 | | | |
| Mg | 0.026 | | | | 0.121 | |
| Ca | | | | | | |
| Na | 0.018 | 1.818 | 0.084 | 1.864 | 0.171 | 1.926 |
| K | 1.800 | | 1.780 | | 1.755 | |
| F | 2.909 | | 3.082 | | 0.140 | |
| OH | n.d. | | n.d. | | n.d. | |

n.d. - not determined

Table 2 Some microprobe analyses of micas from the non-megacrystic areas of the Austell granite, with cell formulae.

# Discussion and Conclusions

Preliminary results from experiments conducted so far show that the behaviour of Li and F in granitic melts is very different. The lowering of liquidus temperatures suggest that Li acts simply as a flux by depolymerizing the melt structure and that the type of complexing postulated by Manning with the addition of F does not occur with the addition of Li. Further, compositions containing 4 wt % $Li_2O$ or more can crystallize petalite as the primary liquidus phase at temperatures as high as 760°C. This implies that petalite in the Meldon aplite could have formed directly from the melt. Petalite does not occur in the St Austell granite because there is simply not enough $Li_2O$ present.

Whole rock analyses show that high $Li_2O$ values are accompanied by high F and $P_2O_5$ values. It is likely that some of the $P_2O_5$ may be present as amblygonite. These analyses also show that while Li and F are concentrated by magmatic processes Li is lost when the rock undergoes hydrothermal alteration, while F may be concentrated in the form of fluorite. The mica analyses show the same effect, the zinnwaldites being replaced by muscovites containing little $Li_2O$ and F, the $Li_2O$ is assumed to have been lost but the F becomes concentrated as fluorite, often along the cleavages of the new mica.

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# Low-grade, regional metamorphism across the Perranporth-Pentewan Line, Cornwall.

T.J. PRIMMER

Primmer, T.J. 1983. Low-grade regional metamorphism across the Perranporth-Pentewan Line, Cornwall. *Proc. Ussher Soc., 5,* 421427.

X-ray diffraction analysis of Devonian slates north and south of the Perranporth-Pentewan line has established that two different regional metamorphic environments exist. To the south of the line, illite crystallinity and $b_o$ determinations of illite/phengite minerals characterize an increasing P/T environment which may be related to comparable data obtained from rocks in the Start complex. North of the line, an essentially low pressure, epizonal (low-greenschist) metamorphic environment with later readjustment of the $b_o$ parameter may be associated with a large D2 backfold development which extends from west Cornwall to south Devon. The juxtaposition of the different metamorphic environments along the Perranporth-Pentewan line is thus related to the Start boundary. This information emphasizes that no link between the Start boundary and the Dodman or Lizard areas exists.

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## Introduction

The Perranporth-Pentewan line has long been recognised as a major tectonic and stratigraphic discontinuity within the Upper Palaeozoic of Cornwall. To the north of the line rocks of Lower Devonian age, the Dartmouth and Meadfoot Beds, have been juxtaposed against the Gramscatho Beds, thought to be of Middle Devonian age (Hendriks *et al.,* 1971). Early descriptions of the boundary were confined to the contact between the units at Pentewan. The Bodmin and St Austell sheet memoir (Ussher *et al.,* 1909) describes the exposure in the cliff section at Gamas Head in terms of an unconformity, which has been normally faulted with a northerly hade of $50^{\circ}$· In the Mevagissey memoir (Reid 1907), the section is considered much too confused and contorted for it to be possible to determine the stratigraphic relationships in detail

In a more recent review of the structure in the Perranporth area (Sanderson 1971) it is suggested that the contact between the two series of rocks is a northerly directed slide, separating early isoclinal folds in the south from more open folds in the north, and that the contact has been subsequently folded by a F2 monocline with an ESE trend. Henley (1973) produced an alternative explanation in which he considered the structure to be essentially a major fault which has reorientated both F1 and F2 folds. The purpose of this study is to evaluate the extent of low-grade metamorphism in the area and determine whether any significant change in grade occurs across the Perranporth-Pentewan line.

## Parameters used to characterize low-grade metamorphism

It has long been recognized that the crystallinity of the illite/phengite series of clay minerals, as measured by X-ray diffraction techniques, can be used as an indicator of advanced burial diagenesis and of incipient regional metamorphism or 'anchimetamorphism' (Kubler, 1967). The degree of thermal alteration can be measured from the crystallinity of the illite which is reflected in the sharpness of the 001 diffraction peak at 10Å. Two indices have been developed.

(a) The 'sharpness ratio' or Weaver index (Weaver, 1960). This is the ratio of the peak minus background intensities at the 10Å and 10.5Å positions.

(b) The 'peak-width' or Kubler index (Kubler. 1967, 1968). This index refers to the peak-width at half peak height above background and in this study is measured in terms of 'degrees $2\theta$' for CuK$\alpha$radiation (Brazier *et al.* 1979).

Boundaries between the zones mentioned above and higher grades of metamorphism have been refined in some detail by Stalder (1979) and Kisch (1980a,b), by comparison with other low grade metamorphic indicators (e.g. metamorphosed basic igneous rock assemblages, coal rank data, fluid inclusion data).

Geobarometers in low-grade metamorphism are rare except in the case of high pressure glaucophanitic bearing assemblages. A method of empirical geobaric analysis has been developed in pelitic terrains (Sassi and Scolari

1974) based on the position of the 060 diffraction peak (i.e. the $b_O$ lattice spacing) of potassic white micas. Parian *et al.* (1982) have recently extended the study into the analysis of low-grade illite/phengite bearing assemblages. The $b_O$ parameter is a measure of the celadonite content of the mica, which increases with replacement of A1 by Fe and Mg. This relationship appears pressure dependent if temperature and bulk composition of the rock are constant (Ernst, 1963). From this Sassi and Scolari (1974) devised a series of pressure 'facies' based on different P/T terrains, with which the present study can be compared.

## Experimental procedure

Illite crystallinity determinations were made on samples of clay-size particles, prepared by ultrasonic disaggregation with the $<2\mu$ clay fraction separated by a centrifuge and filtration technique. Orientated samples were prepared by smearing the clay fraction onto a glass slide, $b_O$ determinations were made on mechanically powdered samples packed into a modified aluminium cavity mount (Robinson, 1981) in order to enhance 0k0 reflections.

All determinations were done on a Phillips PW1730 diffractometer using Ni-filtered CuKα radiation. Divergence and scatter slits of $1°$ and a receiving slit of 0.1mm were also used. Crystallinity determinations were made from scans at 40kV, 30mA over the two-theta range 7.5-10° at 1/2° 2θ/min. Weaver indices Were determined by fixed time counts (30 secs.) at the appropriate peak (10 and 10.5Å) and background positions. A rapid method of monitoring machine drift within a sample run was by running an 'in house' polished slate standard for its Weaver index ten times before and after the start of a run of samples, Measurement of the Weaver index provides a convenient method of analysing variations in diffraction intensity. Measurements of the position of the 060 peak were made on scans at 40kV, 40mA over the range 58-63° at 1/40 2θ/min. using the quartz 211 peak (1.541Å) as an internal standard.

## Results

40 samples were collected north of the line from the Lower Devonian Dartmouth and Meadfoot Beds and 30 samples from south of the line in the Gramscatho Beds (Fig. l) Initially the samples were analysed for their 'whole-rock'$<2\mu$ size fraction clay mineralogy. In all samples this was found to be composed of varying proportions of illite/phengite and chlorite, presumably formed during the main D 1 phase of deformation which produced the penetrative cleavage in the slates.

The illite crystallinity results of both sets of samples can be represented in terms of a plot of the log of the calculated Weaver index against the measured peak width (Fig. 2). A linear trend would suggest that the two indices could be correlated, however, individual crystallinity values that deviate from the linear trend
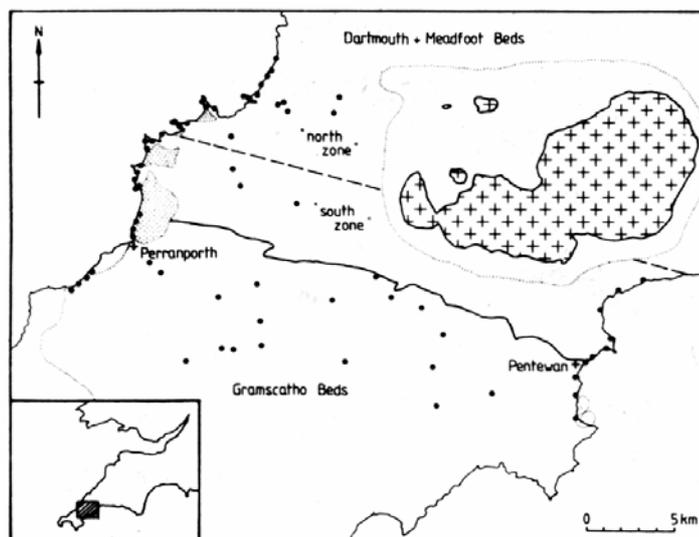


Figure 1. General location map for study area. Solid line represents the Perranporth-Pentewan line; fine stipple, wind blown sand cover; + granite outcrop; dotted line, extent of thermal aureoles around granite.

should be treated with caution. Some samples, especially in the anchizone field have anomalous Weaver values of lower grade than that indicated by peak-width measurements. This appears to be due to the broadening of the base of the 10Å illite peak in these samples; a situation that will affect measurement of the Weaver index, but not the peak-width. The deviation probably represents some characteristic of weathering or structural alteration of the illite. In view of this increased uncertainty in some Weaver values (Table 1) peak-width measurements are used throughout this paper to indicate the degree of illite crystallinity. It can be seen in both graphs that the results straddle the boundary between the anchizone and epizone, the latter being true greenschist metamorphism (Kubler 1967). Work on metamorphic assemblages in basic igneous rocks in the Padstow area to the north (Floyd and Rowbotham 1982) and in the Roseland area to the south (Barnes and Andrews 1981), suggest predominantly lower grades of metamorphism; i.e. prehnite-pumpellyite and pumpellyite-actinolite facies, which should reflect wholly anchizone crystallinity values. It appears therefore, that on a regional scale there exists a metamorphic 'high' along the Perranporth-Pentewan line. Similarly, calculated $b_O$ parameters were treated in a statistical fashion in terms of a cumulative frequency curve for each group of samples or 'population', (Fig. 3). Sassi and Scolari (1974) represented their reference 'facies series` the same way. Figure 3 shows that the populations from north and south of the line both plot between the reference curves of Bosost (Central Pyrenees) and N. New Hampshire indicating that a predominantly low pressure metamorphic environment existed throughout the region. Bearing in mind that the curves display a statistical distribution of $b_O$ values, it should be noted
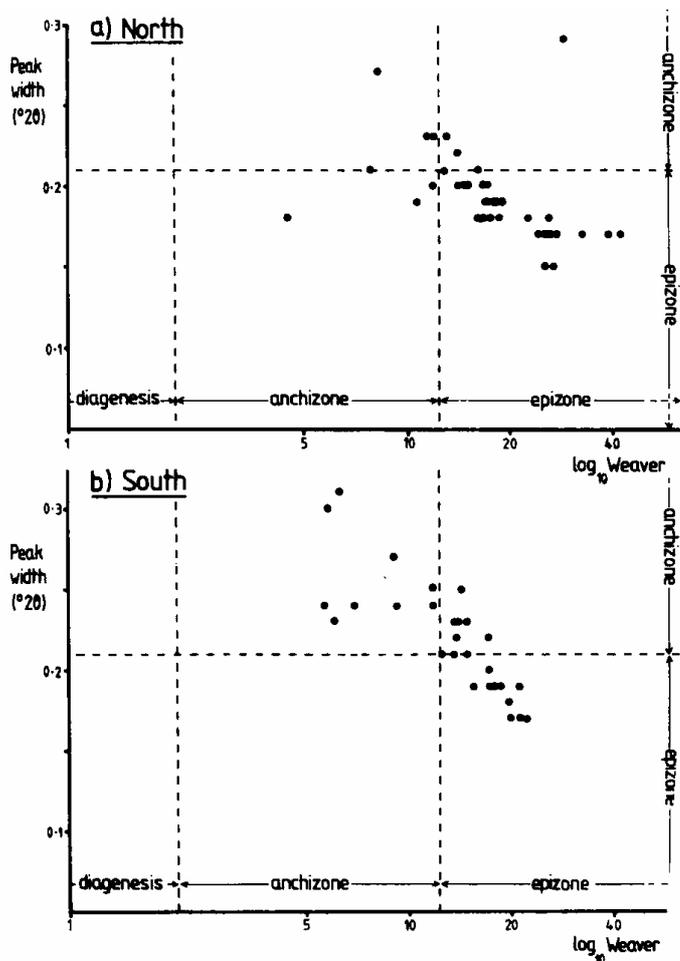
Figure 2. Illite crystallinity values, plotted as peak width (°2θ) against $\log_{10}$ Weaver index. Metamorphic zone boundaries after Brazier *et al.,* (1979) and Kisch (1980a).
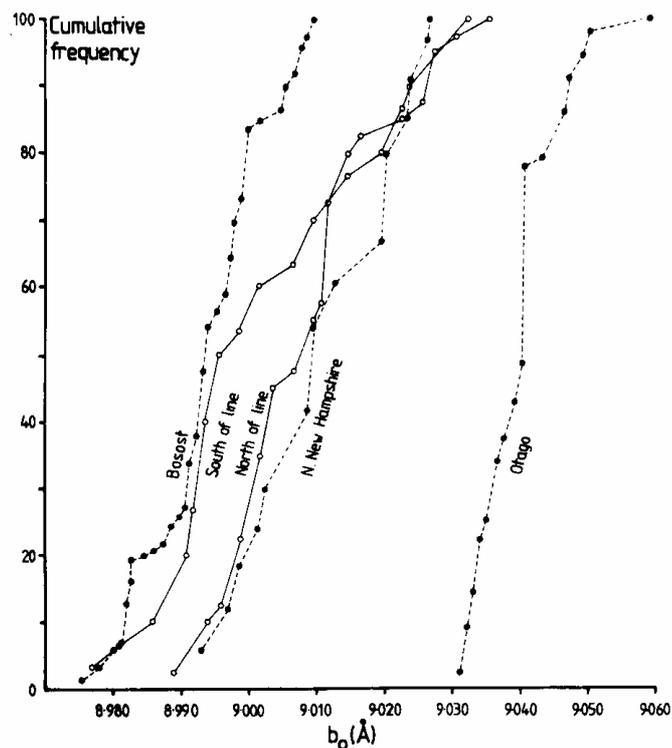


Figure 3. bo cumulative curves for samples from study area compared with reference curves from Bosost (Central Pyrenees), N. New Hampshire (U.S.A.) and Otago (New Zealand), after Sassi and Scolari (1974).

that both curves from the study area, especially the curve representing the population south of the line, may have a somewhat bimodal distribution, shown by the change in gradient in the upper part of the frequency curve (they can be seen to cross over). This suggests that each curve may be composed of two smaller, yet distinct populations.

To consider how illite crystallinity and $b_o$ values are related over the region studied, trends in results in terms of two coastal traverses along the NW and SE coasts can be considered (Fig. 4). The degree of crystallinity appears quite consistent across the line from the north (L. Devonian) to the south {Gramscatho} on the NW coast. However the $b_o$ value decreases in the L. Devonian from the north to the line and shows a similar trend in the Gramscatho to the south with some degree of change across the line. In the Gramscatho series it appears than an increase in crystallinity (i.e. a decrease in peak-width) is accompanied by an increase in the $b_o$ parameter. A similar relationship may be seen on the SE coast in the Gramscatho series. North of Pentewan, no such agreement occurs, in fact the reverse is seen with a slight increase in metamorphic grade being associated with a reduction in the $b_O$ parameter.

It now appears possible to resolve the distinct $b_O$ populations hinted at earlier:

(a) North of the Perranporth-Pentewan line, the samples collected can be divided into two smaller zones which give rise to two distinct populations (Fig. 5). The 'north zone' covers an area from the north to a line extending ESE from Newquay to St Austell Bay (Fig. 1). The 'south zone' covers the area south of this to the Perranporth-Pentewan line. Table 1 shows that there is a measurable difference in mean $b_O$ values with little difference in little crystallinity.

(b) South of the Perranporth-Pentewan line, there is a relationship between an increase in $b_O$ values and an increase in illite crystallinity. Thus, if the samples giving peak width values 0.21° 2 , i.e. anchizone, are separated from the samples giving peak width values 0.21° 2 , i.e. epizone; it is possible to obtain two distinct $b_O$ populations (Fig. 5), the mean $b_O$ values of which can be compared on Table 1.
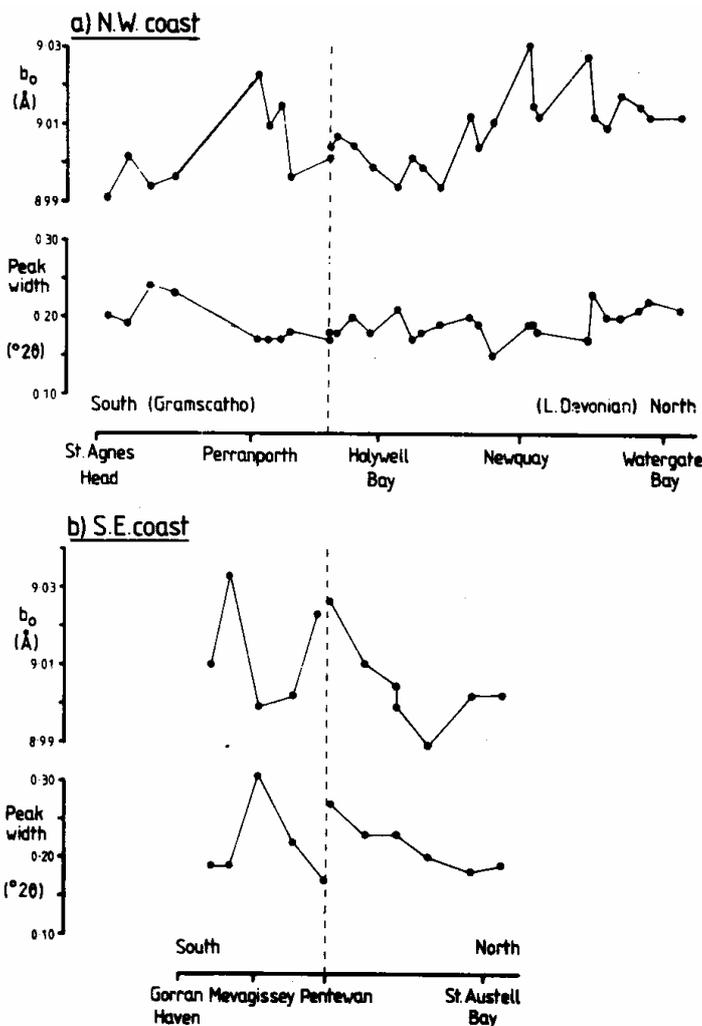
Figure 4. Illite crystallinity (peak width values) and $b_o$ values in terms of 2 coastal traverses. (a) N. W. coast (b) S.E. coast. Dotted line represents the position of outcrop of the Perranporth-Pentewan line on each traverse.



Figure 5. $b_o$ cumulative curves for samples in study area grouped: (a) north of the line according to location (see Fig. 1), (b) south of the line according to metamorphic grade.

## Interpretation of Results

Authors in the past have attempted to link the Perranporth-Pentewan line eastwards with the Start boundary on structural and stratigraphic grounds. (Dearman, 1971; Sanderson and Dearman, 1973; Sadler, 1974; Matthews, 1977). Sadler in particular has strongly questioned the idea of a continuous Lizard-Dodman-Start thrust as described by Hendriks (1939). More recently X-ray diffraction work on rocks juxtaposed at the Start boundary (Robinson, 1981) has provided data which can be compared with that produced here (Table 1). The first direct comparison that can be made is the similarity in $b_o$ values between samples from the Lower Devonian outcrop north of the Start boundary and those from the 'south zone' directly to the north of the Perranporth-Pentewan line. Robinson mentions that the South Devon rocks have Weaver indices in excess of 20 and are thus of comparable metamorphic grade to those

in Cornwall. There is however a marked difference in $b_o$ values for the Gramscatho series and the Start rocks (Table 1), although as explained above an increase in $t_o$ values matches an increase in illite crystallinity. Robinson (1981) presents no crystallinity data on the Start rocks, but petrographic work on the greenschists within the series (Tilley, 1923) suggests that at least low-grades of metamorphism were reached. Therefore, as the Start rocks have been metamorphosed to a higher grade than the Gramscatho series, they might reasonably be thought to have obtained a higher mean $b_o$ value if the two units are to be linked in any way. The relationship between the Start data and that produced for the Gramscatho Beds are shown in Figure 6. It implies that it is possible to see the result of a variable pressure/temperature regime through a low-grade metamorphic terrain associating the Start complex with the Gramscatho series south of the Perranporth-Pentewan line.

If it is possible to relate these two areas in such a fashion, the break in mean $b_o$ values seen across the Start boundary should also be recorded across the Perranporth - Pentewan line. However, as the

Table 1. Mean illite crystallinity and $b_o$ values from study area. Data for S. Devon and Start Schists from Robinson (1981).

North of Perranporth - Pentewan line

|  | Peak-width (°2θ) | | Weaver index | | bo (Å) | | |
|---|---|---|---|---|---|---|---|
|  | $\bar{x}$ | σ | $\bar{x}$ | σ | $\bar{x}$ | σ | N |
| North Zone | 0.19 | 0.03 | 17.35 | 6.21 | 9.017 | 0.009 | 20 |
| South Zone | 0.20 | 0.03 | 21.22 | 9.45 | 9.000 | 0.009 | 20 |

South of Perranporth - Pentewan line

|  | Peak-width (°2θ) | | Weaver index | | bo (Å) | | |
|---|---|---|---|---|---|---|---|
|  | $\bar{x}$ | σ | $\bar{x}$ | σ | $\bar{x}$ | σ | N |
| Epizone | 0.18 | 0.01 | 19.04 | 2.13 | 9.011 | 0.015 | 12 |
| Anchizone | 0.24 | 0.03 | 11.51 | 4.00 | 8.998 | 0.014 | 18 |

|  | Weaver Index | bo (Å) | | |
|---|---|---|---|---|
|  | $\bar{x}$ | $\bar{x}$ | σ | N |
| L. Devonian phyllites (S. Devon) | 20 | 9.002 | 0.009 | 13 |
| Start schists | - | 9.032 | 0.019 | 15 |

Gramscatho series have not reached such high grades of metamorphism, their mean $b_o$ values are not as high as that for the Start complex;

$$\text{i.e., } \bar{b_o} \text{ (Start rocks)} - \bar{b_o} \text{ (L. Devonian, S. Devon)} = 0.030\underline{\text{Å}}$$
$$\bar{b_o} \text{ (Gramscatho epizone)} - \bar{b_o} \text{ (L. Devonian, S. zone)} = 0.011\text{Å}$$

but, as can be seen, a recordable break in mean $b_o$ values can be traced across the boundary. It is useful to note here that the Gramscatho $b_o$ values show a high standard deviation, a factor noticed by Robinson in the Start data. He attributed it to compositional variation within the samples collected. Like the Start rocks, the Gramscatho Beds also contain abundant quartz and chlorite which, according to Sassi and Scolari (1974), result in higher $b_o$ values in some samples. As mentioned above the use of the $b_o$ parameter as a discriminant function depends on the consistency of temperature and bulk composition. On the basis of the XRD data presented, the results can be. interpreted as above, however, variance in the $b_o$ values recorded may reflect variations in the bulk composition of the rocks. Only an extensive study of the whole-rock chemistry of the units could account for any such effects and require reinterpretation of the results described.

The north and south zone results from north of the Perranporth-Pentewan line are also represented on Figure 6. Clearly this terrain is different from that south of the line and a simple pressure/temperature regime cannot be inferred. However the different $b_o$ populations correspond well with the division between the structural zones 7 and 8 of Sanderson and Dearman (1973). The transition from recumbent north facing F 1 folds, to folds with inclined F1 axial planes and an associated increase

in the intensity of F2 folding with crenulation cleavage is accompanied by a decrease in the $b_O$ parameter in the slates. The decrease in $b_O$ values also coincides with a change in the K-Ar ages calculated by Dodson and Rex (1971) over the same area. The younger 310-290 m.y. dates being associated with lower $b_O$ values in the 'south zone' and 340-320 m.y. dates corresponding with higher $b_O$ values from the 'north zone'. In constructing a cross-section through S.E. Devon, Shackleton *et al.* (1982) suggested that the southward steepening of the cleavage and associated crenulation cleavage from Torbay to Start Point is related to a large backfold which can be traced across Cornwall to the west. The upright cleavage is believed to be the same as that to the north folded around the backfold. Considering the similarity in $b_O$ values adjacent to the Start boundary and the Perranporth-Pentewan line it seems reasonable to suggest that the decrease in $b_O$ values from north to south is due to a major D2 event modifying the attitude and crenulating the F1 cleavage near the southern margin of the Lower Devonian outcrop.

In measuring the $b_O$ parameter as described above, the technique relies on obtaining an 'average' $b_O$ value for all the illite/phengite in the sample. If there is any modification of the first formed mineral, by either chemical alteration or by later growth of a second mineral phase of a slightly different chemical composition, the XRD technique cannot separate the events. However, it is possible to detect some modification of the 'average' celadonite content in the illite/phengite of these rocks by a major D2 event of regional extent.
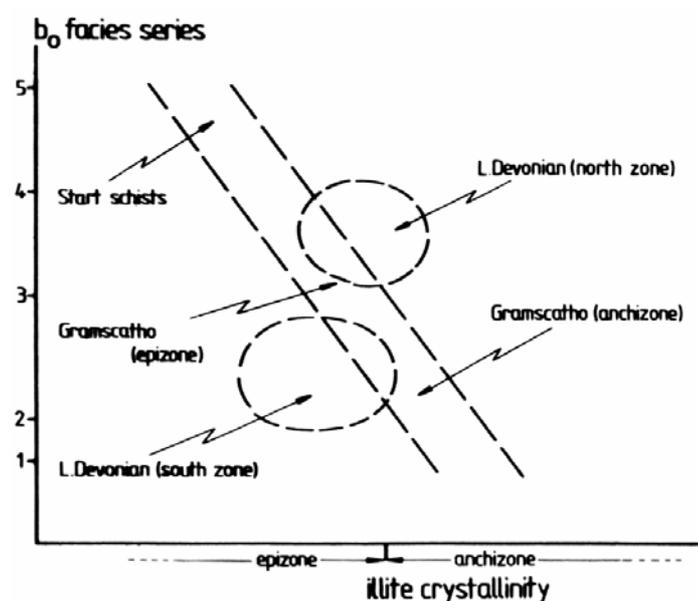


Figure 6. Diagrammatic representation of the rate of change of bo (in terms of Sassi and Scolari's (1974) "fades series" as a function of metamorphic grade (in terms of illite crystallinity).

# Suggested sequence of metamorphic and tectonic events

Assembling all this information it is possible to outline a sequence of metamorphic events related to the tectonic history of the area:

1. North-south compression of the Gramscatho basin (including the Start rocks) resulted in low temperature metamorphism at low to moderate pressures (Barrovian-type metamorphism), transforming the common sedimentary assemblage chlorite + muscovite + K-feldspar + quartz into a metamorphic assemblage, chlorite + illite/phengite + quartz, the new chlorite produced being more aluminous and poorer in Fe, Mg and Si (Velde, 1965). This allows the progressive celadonitization of illite during its crystallization with increasing pressure and temperature.

2. Northward movement superimposed the Gramscatho Beds onto the Meadfoot and Dartmouth Beds. With this northerly migration of deformation, the increasingly 'thin-skinned' nature of the deformation (Isaac *et al.* 1982) precludes the establishment of any significant pressure/temperature gradient. Isaac *et al.* suggest that the allochthon thrust to the north in central Devon and Cornwall never exceeded a thickness of lkm, implying metamorphism was confined to low temperatures and pressures, as indicated by crystallinity and $b_o$ values in the north zone of the L. Devonian. Sanderson (1982)also suggests that there may have been some dextral shear movement on this and other E-W trending structural lines during N-S compression.

3. Continued northward migration of primary deformation with the development of thrust sheets was associated with the development of a large backfold and associated southerly directed thrust movements (Shackleton *et al.* 1982). The secondary deformation superimposed on the D1 fabric may have allowed the decrease in celadonite content with gradual change from fine grained illite to a coarser grained recrystallized phengite during D2. Such transitions to celadonite-poor illite/phengite at low temperatures have been noted in the Salton Sea Geothermal Field (McDowell and Elders, 1980).

4. Later normal faulting during the backfolding downfaulted the Lower Devonian rocks into their present juxtaposition with the Gramscatho and Start rocks. The amount of throw may increase from west to east, bringing higher grade rocks into contact at the Start boundary. In consequence it is emphasized that no link between the Start boundary and the Dodman or Lizard areas can be drawn.

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# Some phosphate minerals at the Megiliggar Rocks, Cornwall

MAURICE STONE
MICHAEL C. GEORGE

Stone, M. and George, M.C., 1983. Some phosphate minerals at the Megiliggar Rocks, Cornwall. *Proc. Ussher Soc., 5*, 428-431.
Further data on amblygonite and triplite as well as new analyses of apatite are presented together with chemical and X-ray data on newly discovered phosphates at this locality. These are triphylite, alluaudite, natromontebrasite and ludlamite. A sequence of crystallization is deduced from textural relationships and shows that triplite developed before zinnwaldite but that amblygonite postdated this mica. Most of the other minerals are later alteration products.
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## Introduction

Rare phosphate minerals of a type encountered in certain types of complex granite pegmatite occur in the granitic sheets of the roof complex of the Tregonning granite, c 2 Km. WNW of Porthleven (Grid Ref. SW612266). Two of these minerals, amblygonite and triplite, have been described already. The former was described petrographically by Stone (1965) and Exley and Stone (1966) and subsequently confirmed by Stone and George (1978) using X-ray diffraction: the latter was determined by X-ray diffraction at Exeter and subsequently confirmed and analysed by energy-dispersive microanalysis at the British Museum (George, Stone, Fejer and Symes, 1981).

This paper supplements the earlier papers with new analytical and/or X-ray data for amblygonite and triplite together with data for apatite and vivianite. It also presents data on new finds of phosphate minerals from Megiliggar, namely triphylite, probable alluaudite and natromontebrasite, and possible ludlamite. Ideal formulae for all these minerals are given in Table 1.

Analytical data were obtained with the wavelength dispersive Microscan 9 electron microprobe at the Geology Department of the University of Oxford and the energy dispersive Link microanalyser attached to the Philips 50lB scanning electron microscope in the Geology. Department of the University of Exeter. In the analysis shown in Table 2, col. 1, F was determined using an automated selective ion electrode. X-ray diffraction data were obtained by conventional powder photography using a 114.6mm diameter Debye-Scherrer camera mounted on a Philips PW1010/80 generator, employing both Fe and Co radiation.

## Phosphate minerals

### Apatite

A complete analysis from a roof pegmatite (MS0810) is given in Table 2, col. 1. H20+ was not determined and is

Table 1. Formulae of phosphate minerals described in text.

| | |
|---|---|
| Fluorapatite | $Ca_5(PO_4)_3F$ |
| Amblygonite | $(Li,Na) Al (PO_4) (F,OH)$ |
| Triplite | $(Mn,Fe)_2PO_4(F,OH)$ |
| Triphylite | $LiFePO_4$ |
| Lithiophylite | $LiMnPO_4$ |
| Alluaudite | $Na(Fe,Mn)_3(PO_4)_3$ |
| Natromontebrasite | $(Na,Li)Al(PO_4)(OH,F)$ |
| Vivianite | $Fe_3(PO_4)_2.8H_2O$ |
| Ludlamite | $Fe_3(PO_4)_z.4H_2O$ |
| Graftonite | $(Fe,Mn,Ca)_3(PO_4)_2$ |

assumed to be     nil in the calculation of the formula, an assumption justified by the good stoichiometry. The high F content is consistent with the high F contents of other (F,OH) minerals in these rocks and the overall high F contents of the rocks themselves (Stone and George, 1978). An analysis of the same material on the Oxford microprobe gives comparable results (Table 2, col. 2). MnO and tFeO (total iron as FeO) are high for fluorapatites. Single analyses of apatite from a roof aplite (MS0811) and the immediately adjacent contact hornfels are shown in cols. 3 and 4 of Table 2. Note the high MnO content of the apatite from the aplite (itself rich in MnO) compared with that from the hornfels, but the higher tFeO content of the hornfels apatite. The highest MnO content found in apatites from these rocks occurs in material intergrown with triplite (specimen MS0086) as shown in col. 5. Formulae have been calculated on the basis of 12.5 oxygens, except in col. 1 where the formula is based upon 13(O,F).

### Amblygonite

Stone and George (1978) give cell parameters for amblygonite from the roof complex indicating that it is F-rich, perhaps close to the high F end member of the montebrasite-amblygonite series. The Li-rich nature of the mineral is deduced from the high Li content of the rock. Analyses shown in Table 3, cols. 6 and 7, are

Table 2. Analyses of phosphate minerals, Megiliggar Rocks

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | | APATITE | | | | AMBLYGONITE | | | TRIPLITE | | |
| Al2O3 | - | - | - | - | - | 33.68 | 32.89 | - | - | - | | - |
| tFeO | 0.62 | 0.64 | 0.3 | 0.68 | 0.56 | 0.09 | 0.09 | 24.51 | 19.1 | 17.01 | 15.24 | 9.12 |
| MnO | 1.33 | 1.27 | 2.24 | 0.98 | 5.69 | 0.07 | 0.1 | 36.92 | 40.33 | 42.6 | 44.71 | 50.36 |
| MgO | - | 0.06 | 0.04 | 0.04 | n.d. | 0.02 | 0.08 | -0.15 | 0.24 | 0.14 | 0.32 | 0.02 |
| CaO | 53.43 | 53.58 | 52.7 | 55.17 | 49.93 | 0.01 | 0.15 | 2.51 | 3.04 | 3.19 | 1.35 | 0.22 |
| Na2O | - | n.d. | n.d. | n.d. | 0.33 | 3.04 | 0.5 | n.d. | - | 0.66 | 1.34 | 1.46 |
| P2O5 | 41.46 | 39.79 | 39.08 | 37.43 | 40.98 | 43.49 | 49.49 | 31 | 31.16 | 32.98 | 33.67 | 32.82 |
| F | 3.95 | - | - | - | - | - | - | 7.4 | - | - | - | - |
| | 100.79 | 95.34 | 94.36 | 94.3 | 97.49 | 80.40 | 83.30 | 102.34 | 93.87 | 95.92 | 95.29 | 92.52 |
| Less O=F | 1.66 | | | | | | | 3.12 | | | | |
| | 99.13 | | | | | | | 99.22 | | | | |
| others | | | | | | | | 0.34 | | | | |
| | | | | | | | | 99.56 | | | | |

Atoms in Formula

| | 1 | 2 | 3 | 4 | 5 | 6 | 7 | 8 | 9 | 10 | 11 | 12 |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| Al | | | | | | 1.05 | 0.94 | | | | | |
| Fe 2+ | 0.04 | 0.05 | 0.02 | 0.05 | 0.04 | | | 0.77 | 0.6 | 0.51 | 0.46 | 0.28 |
| Mn | 0.1 | 0.09 | 0.17 | 0.07 | 0.41 | | | 1.18 | 1.28 | 1.31 | 1.36 | 1.58 |
| Mg | | 0.01 | 0.01 | 0.01 | | | | | 0.01 | 0.01 | 0.02 | |
| (Σ cations) | 5.01 | 5.16 | 5.19 | 5.42 | 5.09 | 2.05 | 1.94 | 2.05 | 2.01 | 2.00 | 1.98 | 1.98 |
| Ca | 4.87 | 5.01 | 4.99 | 5.29 | 4.59 | | | 0.1 | 0.12 | 0.12 | 0.05 | 0.01 |
| Na | | | | | 0.05 | 0.15 | 0.02 | | | 0.05 | 0.09 | 0.11 |
| Li | | | | | | 0.85 | 0.98 | | | | | |
| P | 2.99 | 2.94 | 2.93 | 2.83 | 2.97 | 0.97 | 1.01 | 0.99 | 0.99 | 1 | 1.02 | 1.03 |
| O | 11.94 | 12.5 | 12.5 | 12.5 | 12.5 | 4.43 | 4.49 | 4.09 (5.00) | 4.5 | 4.5 | 4.5 | 4.5 |
| F | 1.06 | | | | | | | 0.88 | | | | |
| OH | | | | | | | | 0.03 | | | | |
| | Ex | Ox | Ox | Ox | Ex | Ox | Ex | BM | Ox | Ex | Ex | Ex |
| N* | 7 | 5 | 1 | 1 | 7 | 1 | 1 | 10 | 4 | 3 | 1 | 1 |

1 and 2:Apatite from roof pegmatite (MS0810)
3 and 4:Apatite from roof aplite and contact hornfels respectively (MS0811)
    5: Apatite intergrown with triplite (MS0086)
6 and 7:Amblygonite from pegmatite
    8: Triplite from Table 1, George et al., 1981.
9 to 12:Triplite from several specimens (MS0086)
tFeO : Total iron oxides as FeO; -: Not determined; n.d. : Not detected; N* : Number of analyses;
Ex : Analysed at Exeter; Ox : Analysed at Oxford; BM : Analysed at British Museum.

calculated to 4 oxygens with $Li_2O$ added to make Li + Na equal to 1. This gives approximately 4.5 oxygens; indeed, if Na is omitted, 1 $Li_2O$ would be required to make oxygen equal 4.5. The variation in Na is probably real, even for the two points analysed on the same material in the two laboratories. Additional formulae (4 analyses) tie within the ranges:

$Li_{0.93-0.98}$ $Na_{0.01-0.07}$ $Al_{0.93-0.95}$ $P_{1.01-1.03}$ $O_{4.5}$ Inclusion of analysed fluorine would probably lower the rather high P.

*Triplite*
George *et al.* (1981) report cell parameters and a chemical analysis for triplite. The latter is shown in Table 2, col. 8 alongside additional new analyses (cols. 9 to 12) which illustrate considerable variation. Great variability in the values of Fe, Mn and Ca are observed, even within a single crystal. Ranges of values based upon 20 analyses in which Mn - Fe - Ca have been summed to 100 are:

Mn: 56.7 - 84.4; Fe: 37.6 - 15.1; Ca: 0.2 - 6.8

In table 2, the formula in col. 8 is based upon 5 (O,OH,F), but in cols. 9 to 12 the formulae are based upon 4.5 oxygens.

*Vivianite*
George *et al* (1981) report vivianite as a blue patchy rim to some triplite masses. Identification is further confirmed by comparison of cell parameters with published data (Table 4).

The following minerals have not previously been recorded from these rocks.-

*Triphylite*
Microanalysis of material intergrown with triplite yields a formula in good agreement with that of triphylite. The formula gives 3.5 oxygens when calculated to one phosphorus atom: 4.0 oxygens are obtained by adding 0.5 $Li_2O$ (Table 3, col. 1). Subsequent X-ray study of green material occurring with triplite confirmed triphylite and yielded cell parameters that lie within the triphylite-rich part of the triphylite-lithiophylite series (Table 4).

*Alluaudite*
Similar green material associated with triplite gives an X-ray pattern matching alluaudite fairly closely (Table 4). The complexity of alluaudite, with something like 20 end-members (Moore and Ito, 1979), means that a formula can only be derived after obtaining a full chemical analysis which includes the determination of ferrous and ferric iron. Two analyses that appear to be consistent with alluaudite are given in Table 3, cols. 2 and 3. Formulae have been calculated on the basis of 12 oxygens after recalculating all FeO to $Fe_2O_3$ This produces a better summation in the analysis: Clearly, more detailed analytical work is required in order to relate the analytical data to alluaudite with more certainty.

Table 3. Analyses of phosphate minerals new to Megiliggar.

| | 1 | 2 | 3 | 4 | 5 |
|---|---|---|---|---|---|
| | Triphylite | Alluaudite | | Natromontebrasite | |
| $Al_2O_3$ | - | - | - | 34.12 | 29.8 |
| tFeO | 32.27 | 24.68 | 27.87 | - | - |
| $Fe_2O_3$ | - | -27.15 | -30.65 | - | - |
| MnO | 12.93 | 18.12 | 12.56 | - | - |
| MgO | 0.08 | n.d | n.d | - | - |
| CaO | 0.04 | 0.32 | 1.42 | - | - |
| $Na_2O$ | 0.84 | 7.47 | 8.92 | 12.86 | 18.48 |
| $P_2O_5$ | 46.32 | 45.83 | 45.53 | 49.75 | 46.99 |
| Others | - | 0.23 | 0.24 | 0.95 | 0.3 |
| | 92.48 | 96.65 | 96.54 | 97.68 | 95.57 |
| | | -99.12 | -99.32 | | |
| Atoms in Formula | | | | | |
| Al | | | | 0.96 | 0.93 |
| Fe2+ | 0.69 | 1.63 | 1.83 | | |
| Mn | 0.28 | 1.22 | 0.84 | | |
| Ca | | 0.03 | 0.12 | | |
| Na | 0.04 | 1.15 | 1.38 | 0.59 | 0.95 |
| Li | 1 | | | 0.4 | |
| P | 1 | 3.09 | 3.06 | 1 | 1.05 |
| O | 4 | 12 | 12 | 4.43 | 4.5 |
| | Ex | Ex | Ex | Ex | Ex |
| N* | 3 | 1 | 1 | 1 | 1 |

1:  Triphylite; 0.5 $Li_2O$ added to formula calculated to 3.5 oxygen atoms.
2:  Alluaudite, calculated to 12 oxygen atoms. Formula $Na_{1-15}(Fe^{3+}_{54}Mn_{41}Ca_{01})_3(P_{1-03}O_4)_3$
3:  Alluaudite, calculated to 12 oxygen atoms. Formula $Na_{1-38}(Fe^{3+}_{61}Mn_{28}Ca_{01})_3(P_{1-02}O_4)_3$
4:  Natromontebrasite; formula calculated to 1.0 phosphorus and adding 0.40 Li and 2.0 oxygen to give 4.43 oxygen atoms.

tFeO: Total iron oxides as FeO; -: Not determined; n.d. - Not detected; N* Number of analyses; Ex Analysed at Exeter.

*Natromontebrasite*
Microanalysis of material closely associated with and apparently an alteration product of amblygonite yielded formulae consistent with natromontebrasite (Table 3, cols. 5 and 6). In col. 5 the data were calculated to 1.0 P atoms and $Li_2O$ added to fill the (Na,Li) site and give c. 4.5 oxygen atoms. On the other hand, a second analysis (col. 6) reveals almost pure natromontebrasite, calculated to 4.5 oxygens. No X-ray pattern for this mineral has yet been obtained.

*Ludlamite*
A partial pattern consistent with that of ludlamite has been noted in the course of the X-ray investigation of vivianite. An equivalent anhydrous formula has been obtained repeatedly by microanalysis: this would agree with the formula for graftonite (Table 1) were it not for the low Ca content of the present data. No X-ray pattern for graftonite has been obtained.

Thus, the presence of triphylite has been established, together with probable alluaudite and natromontebrasite, but the confirmation of ludlamite and identification of other phosphates must await further study.

Table 4. Cell parameters of triphylite, alluaudite and vivianite.

| | Triphylite | | | Alluaudite | | Viviani | |
|---|---|---|---|---|---|---|---|
| | 1 | 2 | 3 | 4 | 5 | 6 | 7 |
| | JCPDS | Megiliggar | JCPDS | JCPDS | Megiliggar | JCPDS | Megiliggar |
| Fe : Mn | 1:0 | 5:2 | 1:1 | | | | |
| a(Å) | 5.98 | 6.05 | 6.06 | 11.03 | 11.13 | 10.06 | 10.10 |
| b(Å) | 10.31 | 10.37 | 10.42 | 12.53 | 12.62 | 13.42 | 13.46 |
| c(Å) | 4.69 | 4.70 | 4.71 | 6.40 | 6.48 | 4.70 | 4.71 |
| β | - | - | - | 97°34' | 97°47' | 104°18' | 104° 34' |

Columns 1,3,4, and 6: Values published in J.C.P.D.S. index.
Columns 2 and 3 : Pegmatite (MS0086)
Column 7 : Pegmatite (MS0086/1)

## Textures

Triphylite examined by microanalysis can be seen on the scanning electron microscope to be intergrown with triplite and occurs as "islands" in the material having the "anhydrous ludlamite" composition. Alluaudite occurs as an alteration product of triphylite and possibly triplite in the polished sections examined in this study. Natromontebrasite is clearly an alteration product of amblygonite. Only triplite, perhaps with triphylite, and amblygonite can be considered with any confidence to be "primary" minerals. These minerals are interstitial to euhedral margins of quartz, potash feldspar and albite. Euhedral quartz within grains of triplite are probably crystals projecting into the triplite perpendicular to the plane of the section. Such features indicate that triplite occurs as cavity fillings. However, zinnwaldite clearly penetrates and includes triplite as well as potash feldspar. Optically continuous islands of triplite occur within large (3mm diameter) grains of zinnwaldite. Tourmaline also fingers and partially encloses triplite. Amblygonite appears to lobe into triplite: it is always interstitial to zinnwaldite, feldspar and quartz and is never included in mica or other minerals.

Thus, it seems that the textural evolution was as follows:

(i) Magmatic growth of quartz and two alkali feldspars (now represented by potash feldspar with exsolved albite and albite with mantles and patches of potash feldspar) followed by some late-stage growth of the potash feldspar and exsolution in the feldspars, possibly coincident with the exsolution of $H_2O$ (and some F) from the last portions of the crystallizing melt. This sequence is based on detailed observations of the textures of lithium-mica granites by the first-named author. It is hoped to publish a full account of the textural evolution of the lithium- mica granites shortly (in press, Proc. Geol. Assoc.).

(ii) Growth of triplite filling some cavities near the completion of crystallization.

(iii) Post-magmatic growth of zinnwaldite and tourmaline.

(iv) Growth of amblygonite interstitial to other minerals and in part or perhaps in whole replacing triplite.

(v) The growth of topaz certainly followed that of both zinnwaldite and tourmaline and preceded the enlargement of quartz grains, but its relationship to the phosphates is uncertain. However, apatite is always included in topaz and never vice versa.

(vi) Hydrothermal alteration processes, possibly in more than one stage, resulted in the production of alluaudite, vivianite, ludlamite (9.) from triplite and/or triphylite and natromontebrasite from amblygonite.

## Summary and Conclusions

Analyses reveal a wide range in Fe and Mn in triplite and confirm many more examples of amblygonite.

Triphylite emerges as a new mineral at this locality confirmed by X-ray diffraction, together with alteration phases such as alluaudite (from triphylite and possibly triplite) and natromontebrasite (from amblygonite). Vivianite and possible ludlamite (and other phosphates) appear to have been derived from triplite, perhaps at a much later stage.

These minerals are common in phosphate-bearing pegmatites elsewhere but have not previously been recognised from this locality,

Examination of textures indicates that "primary" triplite crystallized before zinnwaldite and tourmaline, but "primary" amblygonite post-dates these minerals. Triphylite may be coeval with triplite, but alluaudite and montebrasite are probably hydrothermal products. Vivianite and ludlamite are probably much later and may well have resulted from the weathering of triplite.

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Stone, M. and George, M.C. 1978. Amblygonite in leucogranites of the Tregonning-Godolphin granite, Cornwall. *Mineral. Mag., 42,* 151-152.

# Variscan Thrusting in the Basement of the English Channel and SW Approaches

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J.W.F. EDWARDS

Day, G.A. and Edwards, J.W.F., 1983. Variscan Thrusting in the Basement of the English Channel and SW Approaches. *Proc. Ussher Soc. 5,* 432436.

The origin of linear dipping events in Devonian basement in seismic profiles south and west of the Cornubian Peninsula is discussed, and some are demonstrated to represent thrust planes. The thrusts are north directed and can be correlated with the geology of the Lizard Peninsula. Correlation along strike with thrusts observed in the Start Peninsula is dubious because of late Hercynian dextral movement along NNW directed faults, but no obvious south dipping thrusts are seen in mid-Channel profiles east of the Start Cotentin Ridge. This may be because Permo-Triassic subsidence in the west has increased the angle of dip of the thrusts there so that they can be more easily seen in the seismic profiles, or the Start Cotentin Ridge may represent a major boundary on either side of which the tectonic history has been dissimilar.

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## Introduction

In offshore seismic profiles from the English Channel and Celtic Sea areas prominent events have been observed in the basement rocks underlying sedimentary sequences (Fig. 1). When seen in isolation such features are often ignored, considered to be artificial products of the seismic method, or, if thought real, still ignored because of the difficulty of arriving at a reasonable interpretation. In COCORP data (Brewer and Oliver, 1980) and other land seismic lines (Kenolty *et al,* 1981, Meissner *et al,* 1981) events such as these have been interpreted as thrust planes. The same interpretation has been made of events in marine seismic profiles (Edwards, in press and Leveridge *et al,* in press) and data acquired under the British Institutions' Reflection Profiling Syndicate (BIRPS) programme have indicated the probable existence of major thrusts passing through the upper part of the lithosphere to a depth of 40 kilometres (Smythe *et al*, 1982).

## Validity of seismic events

In offshore seismic lines in this area such events are quite common and it is therefore important to try to establish a basis on which to accept or reject them as representing real geological features. The following possible origins for these events are considered.

1. Artefacts arising from the method of acquisition of field data: identified cases are multiples, side swipe diffraction and reflected refracted returns.

2. Real geological features: the most likely of these are bedding, thrust planes, zones of mineralisation and returns from the tops of buried plutons.

3. There is a possibility that such events might be artificially introduced by some stage of the data processing, but although this might conceivably occur in the processing of a particular survey by one processing house it is thought unlikely to be a general source of spurious events.

In the first category, examples of all of these returns can be found in seismic profiles, however, all but reflected refracted returns are usually easy to recognise and do not constitute more than a small percentage of the events referred to in this paper. The same cannot be said of reflected refracted returns, whose shape depends on the structure of the layer in which the ray is refracted (Fitch 1976). If the layer is planar with a linear horizontal discontinuity causing the reflection, the reflected refracted ray will appear on the seismic section as a linear event whose dip varies, depending on the velocities of the rocks above and forming the refracting layer, and the angle between the direction of the seismic section and the direction of the discontinuity causing the reflection. If the maximum dip of the event is determined from two or more intersecting lines, we show elsewhere (Day and Edwards, in press.) that it will be steeper than one-third s/km for most reflected refracted events and shallower than this for reflections from planes with dips less than about $40^\circ$.

## Origin of events

We have observed events which can be shown by the slope test to be reflections and in one case this has been mapped to near outcrop in the sea floor in the position where, by extrapolating from land observations, one
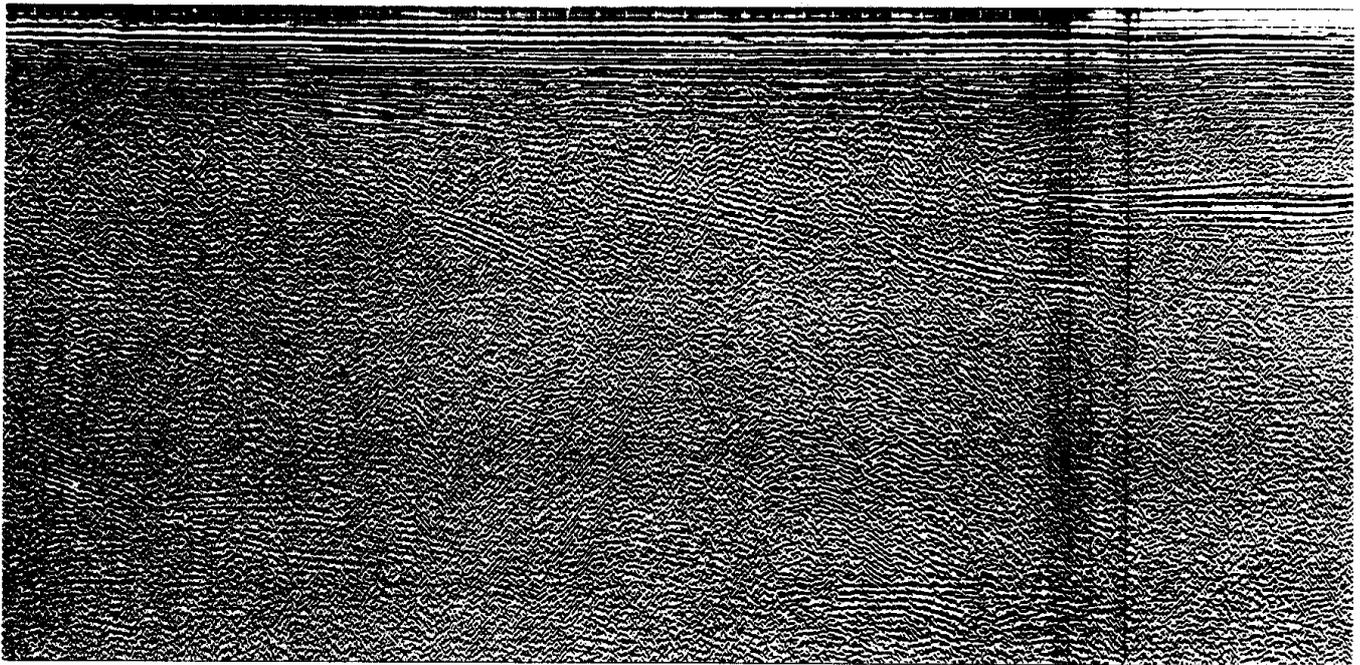
Figure 1(a) Seismic profile running NNW ending about 6km south of Dodman Point.
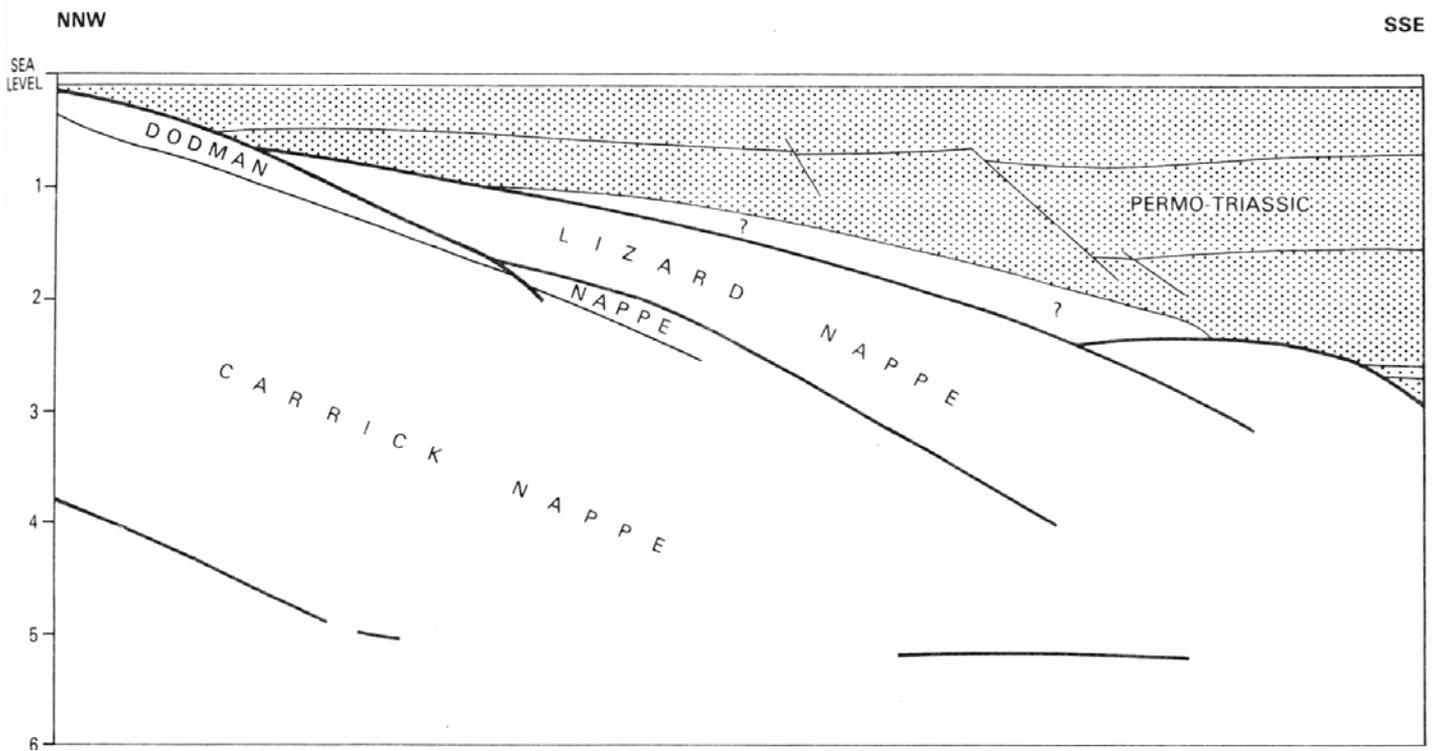


Figure 1(b) Interpretation of (a) showing the strong reflectors in the basement interpreted as thrust planes. Figures on the left side are two-way time in sees. Length of profile is 28km.

would expect to find the sea bed trace of the inferred Carrick Thrust (Leveridge *et al* in press). The seismic event has been mapped to near outcrop both east and west of the Lizard and can be seen dipping in a south southeasterly direction to over five seconds two-way time corresponding to a depth of 13 or 14km below sea level, with a true dip between 30° and 35°· Figure 2 shows the relationship of this event to the inferred position of the Carrick Thrust. It seems probable then that most of the similar deep events seen in seismic profiles from this area, where it can be shown that they are not reflected refractors, arise from thrust planes, although some at least probably arise from the other geological phenomena mentioned earlier: In tile case of zones of mineralisation and the tops and bottoms of buried plutons, these may well be associated with thrust planes and it may be impossible to distinguish them or at least to do more than identify a boundary in the latter instance.
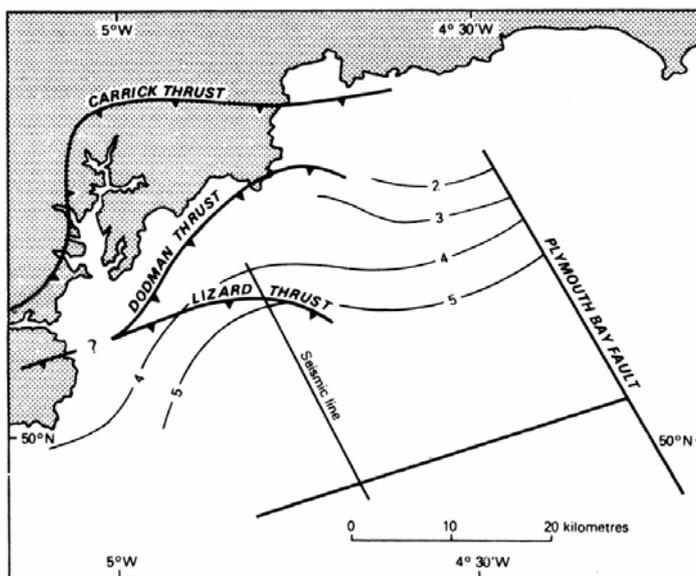


Figure 2. Two-way reflection time map of the prominent reflector in Devonian basement south of Cornwall identified as the Carrick Thrust; isochrons in seconds. Seismic events sub-cropping the Permo-Triassic sediments are shown, which are interpreted as the Dodman and Lizard Thrusts. The position of the Carrick Thrust is inferred from land observations. The line of the seismic section of Figure 1 is shown.

The question of bedding is more complicated. If the mechanism of thrusting consists of detaching whole sections of a formation which then ride up over adjacent sections with bedding planes forming the thrust boundary, (Elliott 1976a) one would expect any bedding in the thrust slabs to be sub-parallel to the thrust planes. Figure 3 shows a seismic profile in which we may be seeing a thrust plane with sub-parallel bedding cut by a reflected refracted event. As far as the dipping events in Plymouth Bay are concerned, we can rule out bedding as a possible origin, for in profiles that extend some distance to the south, they can be seen to be repeated such that if

they were inclined bedding planes, the whole sequence would represent several tens of kilometres of Devonian sediments. This would require the Devonian rocks currently exposed in Cornwall to have been buried to sub-Moho depth.

If we accept then that some at least of these events are caused by dipping thrust planes, the question arises, what is the origin of the reflection? In some cases it is easy to imagine that there is a change of rock type across the boundary. Elliott has described a mechanism in which material being eroded from the exposed toe of a thrust is deposited on the footwall and subsequently over-ridden by the advancing thrust, and in some thrusts there may be dissimilar rocks at depth where similar rock types, or the same rock unit, straddle the thrust at the surface, but in other cases we know that a thrust travels some distance through a relatively homogeneous unit or is contained completely within that unit, and so we need a mechanism which gives rise to a seismic reflection that is in some way connected with the thrust plane itself.
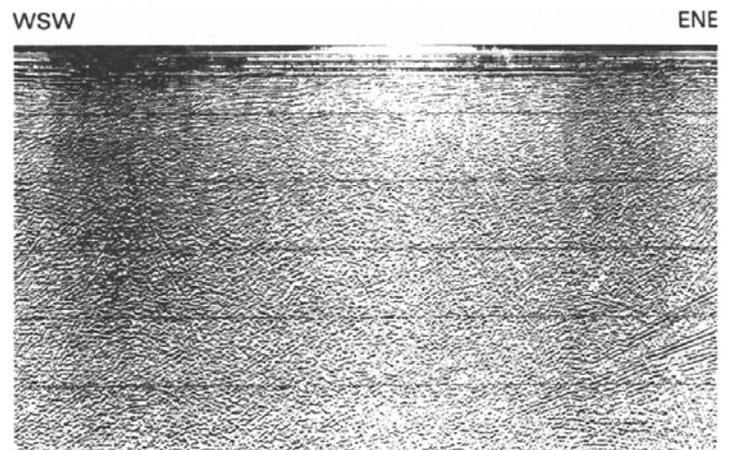


Figure 3(a) Seismic section running ENE towards the southern end of the Start Peninsula (Published by courtesy of Shell UK Exploration and Production Ltd).
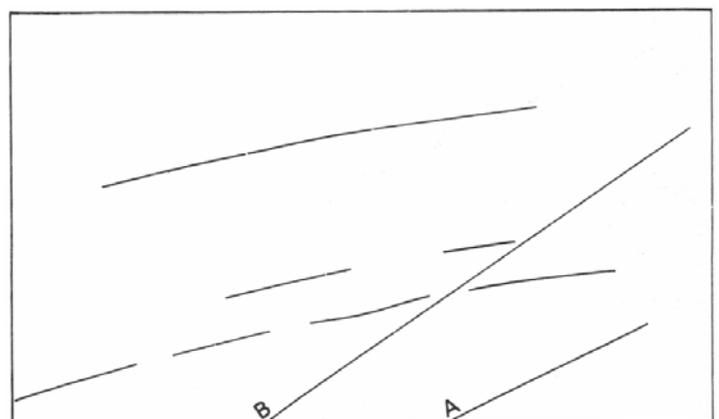


Figure 3(b) Interpretation of (a). A = thrust plane, B= reflected refractor arising from a fault off-section. Other lines are interpreted as bedding within the thrust slice.

This mechanism has to change the acoustic impedance of the rock and could do so by altering its chemical or mechanical properties. An example of the former would be ground water passing preferentially along a thrust plane dissolving out certain constituents and possibly replacing them with others, but the inhomogeneity must occur over a distance normal to the thrust plane which is at least comparable with the wavelength of the acoustic energy being reflected. Unless we are dealing with a multi-layer sandwich of thrust planes, this seems unlikely. Mechanical alteration might conceivably extend some considerable distance from the plane of maximum slippage. Meissner *et al* (1981) point out that high pore pressure and grain diminution, which are often associated with thrust faults, lead to decreases in velocities. In their reflection-refraction work they deduce that their low velocity layer is between 40 and 100m thick. Elliott (1976b) describes a layer of regional penetrative deformation that may be several kilometres thick.

## Extent of thrusting

We are using the technique referred to above to discount reflected refractors as the origin for some of the events seen extensively in seismic profiles from this area. In addition to the area south of the Lizard Peninsula mentioned earlier, there is a zone farther west where an event can be identified in a number of adjacent seismic profiles (Fig. 4). This event, dipping gently southwards, deep within the basement, might be the westward extension of the Carrick Thrust, but we have not so far been able to demonstrate continuity between them. Gravity observations (Edwards, in press) suggest the existence of a NNW-SSE transcurrent fault between these two areas and such a fault would make it difficult to trace the event along strike. In any case it is more difficult to follow the reflectors in the seismic profiles along strike, which is compatible with thrusts having only a limited lateral extent (Elliott 1976b).
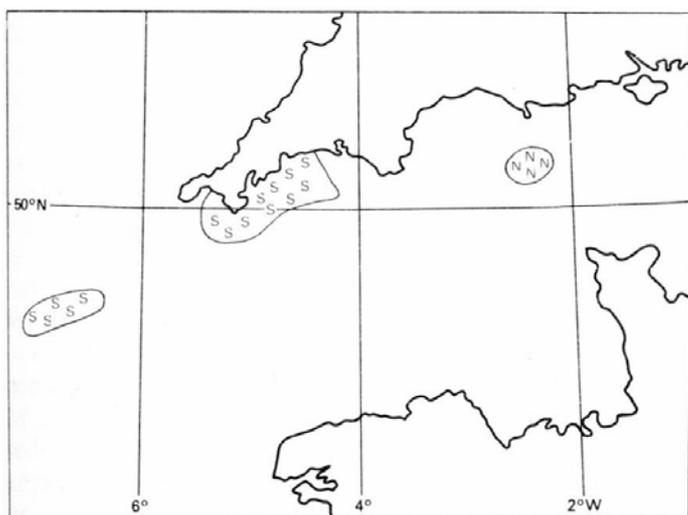


Figure 4. Areas referred to in the text where dipping basement events in seismic sections have been shown to represent reflections from thrust planes. N = north dipping, S = south dipping.

In several seismic lines south of the Cornubian Batholith sub-horizontal events can be seen near the bottom of the section at around six seconds TWT. An example of such an event can be seen in Figure 1 where it would represent a depth of about 14km below sea level. Similar events can be seen under the batholith around the Scilly Isles at 5.8 to 6.0 seconds (Edwards, in press) and under the western part of the Haig Fras Batholith at 4.8 to 5.0 seconds. In both these cases the events are seen only under the southern flanks of the batholith and disappear off the section at either end. A reflector at a depth of 10km beneath North Devon is reported by Mechie (1980) and a refractor is reported by Brooks *et al* (1983) at about 6km beneath the Bristol Channel. We suggest that these sub-horizontal events represent a decollement plane which is gently inclined southwards beneath the Cornubian Peninsula and rises more steeply to form the Variscan Front in South Wales.

The general strike of the thrusts is ENE-WSW and we have tried to find evidence of them in seismic profiles farther east. It is tempting to try to equate them with thrusts described in the Start Peninsula by Coward and McClay (1983) or east of Plymouth by Chapman *et al* (in press) which are directly along strike, but although there are features in the profiles south of the Start Peninsula that might well be thrusts no convincing correlation has so far been made. NNW-SSE trending wrench faults are known to modify the structural geology of South Devon and Cornwall and Dearman (1963) estimated a structural cumulative dextral movement along them of at least 34km. Arthaud and Matte (1977) stressed the importance of dextral strike-slip movements along such faults at the end of the Hercynian orogeny, and we have observed in the seismic profiles a major fault following this trend in Plymouth Bay. It seems likely therefore that any continuation eastwards of the thrust referred to above would now be found nearer the middle of the Channel.

The events in seismic sections from south of the Start Peninsula that appear to be reflections from thrust planes dipping southwards cannot be correlated with those positively identified west of the Plymouth Bay Fault. Farther east no well defined south dipping events have been observed, however northward dipping events have been seen in the basement in an area centred about 40km south of Portland, shown in Figure 4. The most prominent of these events can be tied on several seismic lines and appears to be a real reflection.

The reason for the apparent lack of south dipping thrusts in the central Channel may be that in the Channel we are only recognising these features where they have a pronounced dip in the profile. Seismic profiles indicate over 4secs TWT of Permo-Triassic sediments in Plymouth Bay, so there must have been considerable subsidence after the Hercynian mountain-building episode that produced the thrusting in the Devonian rocks beneath. This implies a tensional regime in Permo-Triassic times, but since the Devonian basement

remained high east of Plymouth Bay, extension must have occurred by sinistral movement principally along the Plymouth Bay Fault. Downwarping of the Devonian basement as this extension took place would have accentuated the southerly dip of the thrusts, but over the Start-Cotentin High these would have remained more nearly horizontal and consequently difficult to recognise in the seismic profiles. The existence of north dipping thrusts south of Portland raises the possibility of there being a major discontinuity crossing the Channel in the Lyme Bay area separating two regions of distinctly different tectonic history.

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# Studies on Fal Estuary Sediment I: Early diagenetic sulphide mineralisation

M.G. THORNE

Thorne, M.G., 1983. Studies on Fal Estuary sediment: I: Early diagenetic sulphide mineralisation. *Proc. Ussher Soc., 5,* 437-444.

There are few documented occurrences of significant quantities of non-ferrous metal sulphides in Recent siliciclastic sediments which do not have volcanic or hydrothermal associations. In this study the Fal Estuary is employed as a natural laboratory to demonstrate that given adequate metal supply non-ferrous sulphides are formed during early diagenesis of estuarine sediment. Copper assays averaging 0.4% Cu are obtained from some estuarine sediments in Restronguet Creek which drains from the abandoned Gwennap copper mines. At the head of this tributary estuary purple bornite and chalcocite are the principal copper minerals and show textures which probably result from the replacement of detrital chalcopyrite. Downstream there is a zoned diagenetic copper sulphide assemblage from orange pink bornite with subordinate chalcopyrite to chalcopyrite with pyrite at the mouth of the estuary. There is no firm evidence of replacement of diagenetic pyrite by copper sulphide.

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## Introduction

Since the historic work of Bastin (1933) and Garlick and Brummer (1951) there has been a steady increase in the number of metal sulphide ore deposits without an obvious igneous or hydrothermal association which are considered to be syndiagenetic with the enclosing sediment. In this paper "syndiagenetic'! is used in the sense of Bissell (1959) to include all stages of sediment formation from transportation to early diagenesis. The famous Cu-Co deposits of Zaire-Zambia and Cu-Pb-Zn deposits of the Kupferschiefer (G.D.R.) are commonly quoted as examples. However few investigations of possible present day natural analogues have been undertaken and interest has largely been concentrated on deposits forming in volcanogenic or hydrothermal environments. For example there is abundant evidence for the presence of non-ferrous metal sulphides in sediments from Red Sea geothermal deeps and reports have been presented by a number of authors including Weber-Diefenbach (1977). Diagenetic recrystallization of these sulphides in sediments from the Atlantis II Deep containing 0.37% CuO has been further discussed by Mossman and Heffernan (1978). In contrast siliciclastic sediments without volcanic or geothermal influences appear to contain non-ferrous metal concentrations one or two orders of magnitude lower than these. For example the most commonly reported mean total copper concentrations for black shales and grey silty shales lie in the range 50-100ppm, Vine and Tourtelot (1970), and various authors quoted in Gad, Cart, and Le Riche

(1969). Similar concentrations were obtained in unconsolidated fjord sediments studied by Presley *et al* (1972) and Krom (1976). Only two reports of the identification of microscopic non-ferrous metal sulphides in sediments without an hydrothermal association are known to the author. Two Australian authors Bubela (1978) and Skyring (1978) both mention Pb4Zn sulphides forming in Spencer Gulf, South Australia downstream from the Port Pirie smelter, in the Annual Reports of the Baas Becking Geobiological Laboratory, Canberra, Australia. However a detailed description of the occurrence is not given. Secondly an occurrence of Zn-Fe sulphide is reported by Luther *et al* (1980) from highly polluted organic rich sediments of a New Jersey, U.S.A. estuary. Thus it is not known whether it is a deficiency in metal supply, fixative sulphide productivity or burial diagenetic mobilization of mineralizing fluids which prevents sulphide mineralization from forming in quantity in Recent siliciclastic sediments. These two companion papers (Part I and Part II) are presented in order to demonstrate that significant quantities of the copper sulphides chalcopyrite and bornite can be formed during deposition and early diagenesis of present day estuarine sediments given an adequate metal supply (Part I) and that on burial these metalliferous sediments might release sufficient dissolved metal for the formation fluids to become potent mineralizing agents for other sediments in the sequence (Part II).

## The Field Areas

The Fal Estuary is situated in south-west Cornwall, forms part of aria coastline and has six major tributaries (Fig. 1). Two of these, Restronguet Creek and the Tresillian River, have been studied to determine their sedimentological, geochemical and mineralogical characteristics. Restronguet Creek is a tributary situated near the mouth of the Fal Estuary which may also be called Carrick Roads. The main central channel in Carrick Roads is 15m deep and as Restronguet Creek is also sheltered from the prevailing south-west wind it is not subjected to the full force of Atlantic gales or tides. Despite this the Spring Tidal range is about 5m and the sediments and structures are dominated by strong flood tidal forces. In the lower estuary most of the sediment is marine derived quartz sand. In the upper estuary, however, vast quantities of silt and fine sand sized copper mining waste from the Gwennap mining district have been admixed to the natural sediments. As a result of this pollution detrital chlorite, tourmaline, feldspar, chalcopyrite, sphalerite, stannite and cassiterite are ubiquitous in Creek sediment. Run off through these polluted sediments is considerably enriched in iron, copper and zinc. As a natural laboratory for investigation of the response of a temperate estuarine environment to high concentrations of dissolved and particulate forms of heavy metals, Restronguet Creek is probably unequalled.
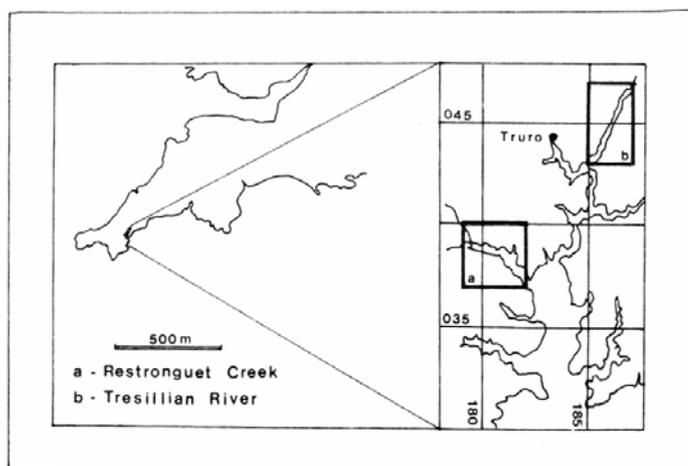


Figure 1. Location map showing Fal Estuary and tributary Restronguet Creek and Tresillian rivers.

## Sedimentology

The important sedimentological features of Restronguet Creek are an extensive creek mouth complex, the intertidal flats and an area of saltmarsh. In many ways these features resemble those which occur in the estuaries on the eastern seaboard of the U.S.A. which were described in Lauff (1967).

*Creek mouth complex*

On a broad scale the creek mouth complex consists of three flood channels, a flood ramp and an ebb shield. The ramp is extensively megarippled with ripples of amplitude 0.35m and wavelength 6m. At slack water and during ebb tide fine grained sediment and organic debris collects in the megaripple troughs and in Places semi permanent braided drainage systems have developed. These features are preserved in the subsurface as planar cross stratified coarse and medium sands with mudstreaks and mudflasers. Some mudstreaks show evidence of lenticular bedding, lensoid silty inclusions and trough cross bedding.

The ebb shield is an elongated rampart which has been thrown up at the crest of the flood ramp and directs ebb runoff away from the face of the ramp. The shield is complexly rippled with at least three major interfering sets. For example on the upper ebb shield an early flood oriented set (wavelength lm, amplitude 10cms), is modified by a later linguoid, parasitic ebb set (wavelength 5cms, amplitude 2cms). Where preserved these sediments occur as thinly bedded, irregularly cross stratified fine sands with sporadic thin mudflasers and rare mudstreaks.

The plan, (Fig. 2), shows the location of the creek mouth complex, its divisions and major sediment sources, and a schematic cross section of the complex is presented in (Fig. 3). As may be seen sediments beneath the present day flood ramp and ebb shield sands are bioturbated silts with some fine sands. These seem to be remnants of an older intertidal flat-channel complex which is now being partially destroyed by the shoreward retreating flood ramp. This retreat is partly due to major flood channel reversion but is also due to a decline in sediment supply since mining and tin streaming in the tributary Carrion River ceased.
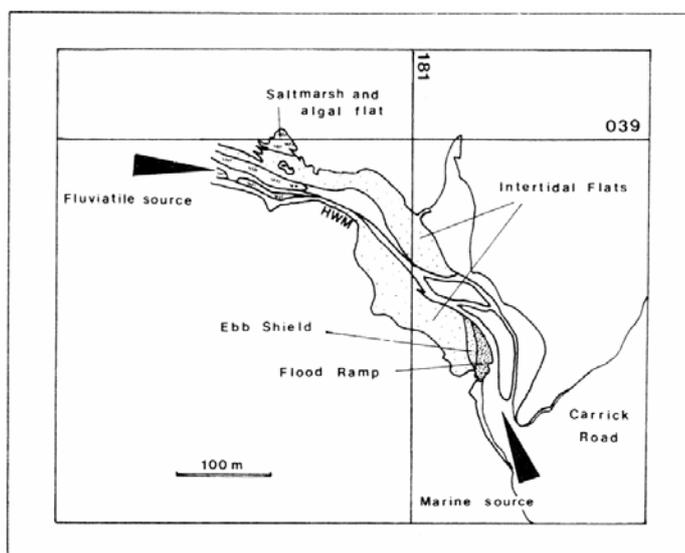


Figure 2. Major sedimentary environments in Restronguet Creek illustrating marine and fluviatile sediment sources.
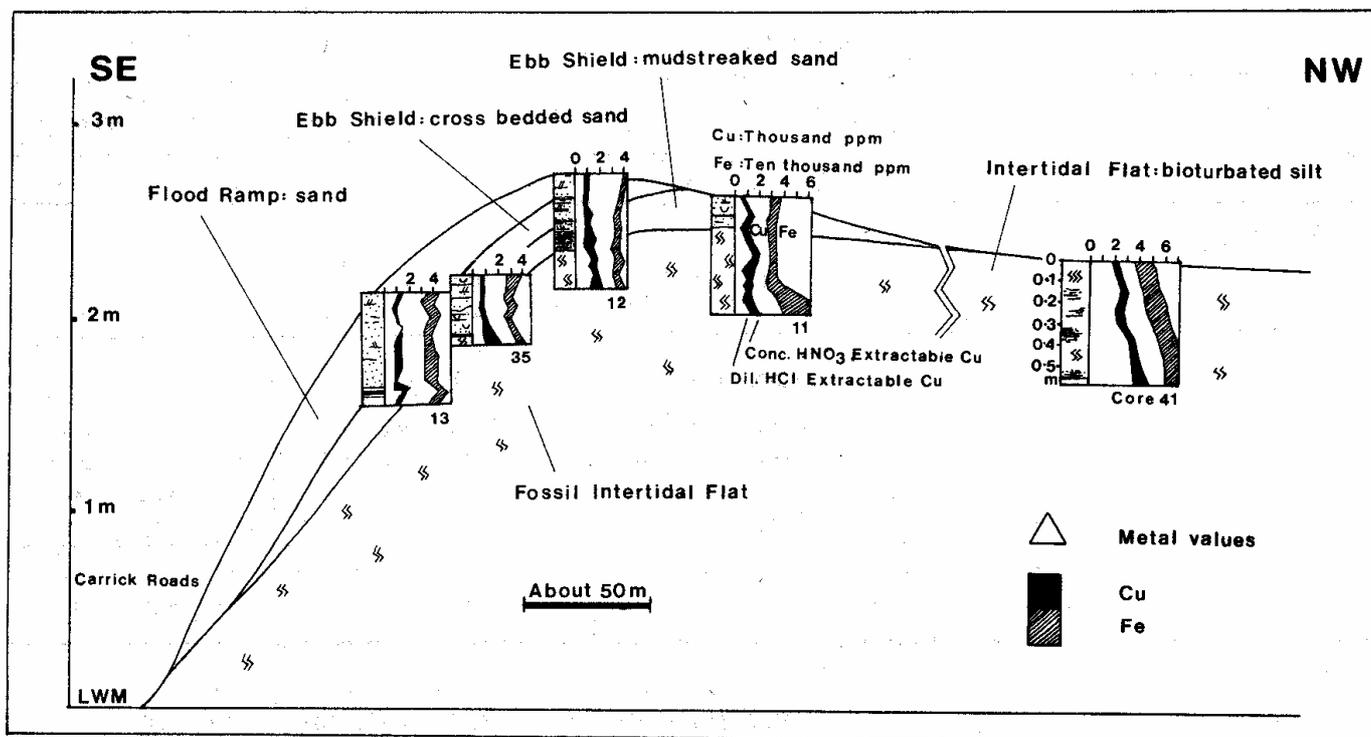
Figure 3. Cross-section of Restronguet Creek mouth complex showing copper and iron concentrations in core samples.

*Intertidal Flats*

Intertidal flats in Restronguet Creek are extensive but exhibit few structures other than crab pits, algal hummocks, and very low amplitude ripple marks adjacent to the ebb shield. The sediments are pervasively worm bioturbated such that evidence in the subsurface of abandoned stream channels and buried algal hummocks is all but obliterated. These features can now be delineated only by a slight coarsening in sediment grain size and an increase in organic carbon content associated with rather vaguely laminated fine silt beds.

As on the flood ramp there is evidence that sediment supply from the Carnon River is much reduced and a linear zone of algal hummocks which was once parallel to a major central channel is now retreating shorewards. As a result, in Core 41, algal sediments are found buried beneath more recent finer grained intertidal flat sediment. (See Fig. 3).

*Saltmarsh*

Saltmarsh in Restronguet Creek is established on areas of elevated, thinly bedded silts which are in part artificially deposited mine tailings. The saltmarsh consists of 12cms of root penetrated silt which is completely oxidized and stained orange by iron oxide, underlain by 12-30cms of reduced sulphurous sediment with decaying organic matter, black amorphous iron sulphide and thin lenses and coatings of jarostic iron sulphate. The salt-marsh is currently being steadily eroded and is retreating shorewards except in very limited areas on the crests of point bars in minor tributary streams.

# Geochemistry and sulphide mineralogy

Samples of Restronguet Creek sediment from pits and vertical cores have been analysed by atomic absorption spectrophotometry for Fe, Mn, Cu, Pb and Zn after partial dissolution firstly in 2 M HC 1 and then separately in conc. $HNO_3$. Orientation experiments showed that 2M HC1 effectively extracts these metals when present as pore water salts, iron and manganese oxide, iron phosphate, iron, copper, zinc, lead and manganese monosulphide, manganese and iron carbonate, chlorite, and also when adsorbed on clay, oxide and organic surfaces.

Treatment with concentrated nitric acid followed by dilute hydrochloric acid, in addition, also decomposes pyrite, chalcopyrite, stannite and organic materials. Thus the difference between assays employing these two dissolution methods, which has been called the △ metal value, should represent iron and copper mostly present in the sediment as pyrite, chalcopyrite and to a lesser extent organic complexes. For other metals the △value most probably broadly indicates only the degree of organic complexation.

A total of 620 samples of all types were analysed and the results have been compiled into Table 1. This shows the mean metal concentrations obtained in subdivisions of major Fal Estuary environments and, for comparison some results from totally unpolluted and unmineralized coarse silt grade inter-tidal flat sediment from Freiston Low, Wash, E. Anglia. Clearly Restronguet Creek has

Table 1. Mean metal concentrations.

| Major Environment | Subdivision | Dil HCl Extractable (ppm) | | | | | Conc. $HNO_3$ Extractable (ppm) | | | | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| | | Fe | Mn | Cu | Pb | Zn | Fe | Mn | Cu | Pb | Zn |
| **Restronguet Creek** Creek Mouth Complex | Flood Ramp | 37469 | 390 | 605 | 215 | 909 | 43605 | 463 | 1109 | 223 | 1053 |
| | Ebb Shield | 33158 | 402 | 688 | 201 | 839 | 43082 | 571 | 1229 | 202 | 1052 |
| Intertidal Flats | Lower Coarse-Grained Flats | 32500 | 414 | 813 | 169 | 867 | 43320 | 492 | 1399 | 181 | 1103 |
| | Lower Flats | 43825 | 563 | 1681 | 284 | 1542 | 58578 | 653 | 2634 | 304 | 1892 |
| | Buried Algal Banks | 51840 | 592 | 2904 | 368 | 2376 | 64680 | 662 | 3686 | 381 | 2861 |
| | Algal Banks | 56123 | 756 | 2507 | 360 | 2024 | 69489 | 871 | 3613 | 374 | 2499 |
| | Upper Flats | 53916 | 676 | 3032 | 423 | 2751 | 66374 | 849 | 4081 | 459 | 3174 |
| Saltmarsh | Oxidized Saltmarsh | 48950 | 339 | 1851 | 337 | 1639 | 60860 | 473 | 2003 | 324 | 1715 |
| | Reduced Saltmarsh | 52000 | 517 | 2860 | 420 | 2300 | 62400 | 650 | 3060 | 425 | 2300 |
| **Tresillian River** Intertidal Flats | | 19622 | 159 | 377 | 172 | 599 | 26965 | 205 | 488 | 186 | 666 |
| Channel Slopes | | 19733 | 182 | 362 | 163 | 550 | 28700 | 229 | 437 | 167 | 619 |
| Point Bars | | 19273 | 131 | 297 | 150 | 610 | 30204 | 218 | 388 | 165 | 718 |
| **Freiston Low, Wash** Intertidal Flats | | 16089 | 316 | 26 | 63 | 107 | 23548 | 380 | 33 | 76 | 119 |

suffered heavy metal pollution on a major scale whilst metal concentrations in the Tresillian River are similar to those obtained in unpolluted streams draining weakly mineralized hinterlands in other areas of south-west England (Table 2, Aston and Thornton, 1975). These conditions are suitable for evaluation of the effect of anomalous metal supply in promoting the development of diagenetic metal sulphides.

Results from representative vertical core samples have already been presented in Figure 3 and show that total metal values are controlled to a great extent by sediment grain size. The highest values of about 6% Fe and almost 0.4% Cu are certainly not restricted to those upper estuarine areas with visible mine tailings pollution. Evidently iron and copper have been mobilized downstream into predominantly marine derived sediments in the creek mouth complex.

The mode of occurrence of these metals cannot be determined geochemically as specific dissolution techniques for oxides, monosulphides and dilute acid soluble silicates (e.g. chlorite) have not been developed. However by combining geochemical evidence with the results of mineralogical and semi-quantitative X.R.D. studies it is possible to suggest very broad trends in mineralogical occurrence at least for iron and copper. Table 2 gives △metal values for major sediment types in

Restronguet Creek and some comparative values for Tresillian River and Wash sediments. When expressed as a percentage of the total nitric acid extractable metal these △ Fe and △ Cu values show systematic trends Table 3. For example in Restronguet Creek iron is largely present as thin oxide coatings to mineral grains on the flood ramp which give a low △Fe/$HNO_3$Fe value of 14%. With increasing fine grained sediment, organic carbon and sulphide contents this ratio rises to a maximum of 25% in lower intertidal flats sediment where pyrite is ubiquitous. The proportion of acid soluble chlorite has also been indicated by semi-quantitative X.R.D. study. Most of this chlorite is detrital but a significant number of samples from the algal banks and upper flats have characteristically different 14Å peak area and 14Å-7Å peak interval statistics. This is tentatively attributed to development of authigenic iron rich chlorite in these areas. On the saltmarsh detrital pyrite is more common and △ Fe/$HNO_3$Fe rises to 24%.

Copper shows somewhat similar zonation in that creek mouth complex sediment is mineralized by chalcopyrite and organically associated copper whereas intertidal flat sediment is mineralized by dilute acid soluble bornite and adsorbed copper. As a result △ Cu percentages decline from 45% on the creek mouth complex to only 21% in organic rich algal bank sediment on the intertidal flats. Saltmarsh sediment contains little chalcopyrite and the

Table 2. Δ Metal values.

| Major Environment | Subdivision | Δ Metal Values (Conc. $HNO_3$ Extractable Metal – dil. HCl Extractable Metal) | | | | |
|---|---|---|---|---|---|---|
| | | Fe | Mn | Cu | Pb | Zn |
| Restronguet Creek Creek Mouth Complex | Flood Ramp | 6136 | 73 | 504 | 8 | 144 |
| | Ebb Shield | 9924 | 169 | 541 | 1 | 213 |
| Intertidal Flats | Lower Coarse-Grained Flats | 10820 | 78 | 586 | 12 | 236 |
| | Lower Flats | 14753 | 90 | 953 | 20 | 350 |
| | Buried Algal Banks | 12840 | 70 | 782 | 13 | 475 |
| | Algal Banks | 13366 | 115 | 1106 | 14 | 475 |
| | Upper Flats | 12458 | 173 | 1049 | 36 | 423 |
| Saltmarsh | Oxidized Saltmarsh | 11910 | 134 | 152 | NIL | 76 |
| | Reduced Saltmarsh | 10400 | 133 | 200 | 5 | NIL |
| Tresillian River Intertidal Flats | | 7343 | 46 | 111 | 14 | 67 |
| Channel Slopes | | 8967 | 47 | 75 | 4 | 69 |
| Point Bars | | 10931 | 87 | 91 | 15 | 108 |
| Freiston Low, Wash Intertidal Flats | | 7459 | 64 | 7 | 13 | 12 |

Table 3. Iron and copper mineralogy.

| Sedimentary Division | $\left(\dfrac{\Delta \text{Fe} \times 100}{\text{Total Fe } HNO_3}\right)$ % | Iron Mineralogy | $\left(\dfrac{\Delta \text{Cu} \times 100}{\text{Total Cu } HNO_3}\right)$ % | Copper Mineralogy |
|---|---|---|---|---|
| **Restronguet Creek Creek Mouth Complex** | | | | |
| Flood ramp | 14 | Iron oxide > pyrite | 45 | Chalcopyrite + organically associated copper > bornite + adsorbed copper. |
| Ebb shield | 23 | Pyrite + monosulphide > iron oxide | 44 | |
| **Intertidal Flats** | | | | |
| Lower flats | 25 | Pyrite + monosulphide iron rich silicate | 39 | Bornite + chalcopyrite + adsorbed copper |
| Algal banks | 20 | Monosulphide + iron rich silicate > pyrite | 21 | Bornite + adsorbed copper > chalcopyrite |
| Buried algal banks | 19 | " | 31 | " |
| Upper flats | 19 | " | 26 | " |
| **Saltmarsh** | | | | |
| Oxidized | 24 | Iron > pyrite | 8 | Bornite + chalcocite + adsorbed copper >> chalcopyrite |
| Reduced | 20 | Monosulphide + iron rich silicate + iron oxide > pyrite | 7 | " |
| **Tresillian River** | | | | |
| Intertidal flats | 37 | Monosulphide + pyrite | 23 | Adsorbed copper + organically associated copper + bornite? >> chalcopyrite |
| Channel slopes | 45 | Pyrite + monosulphide | 17 | |
| Point bars | 57 | Pyrite + monosulphide | 23 | |
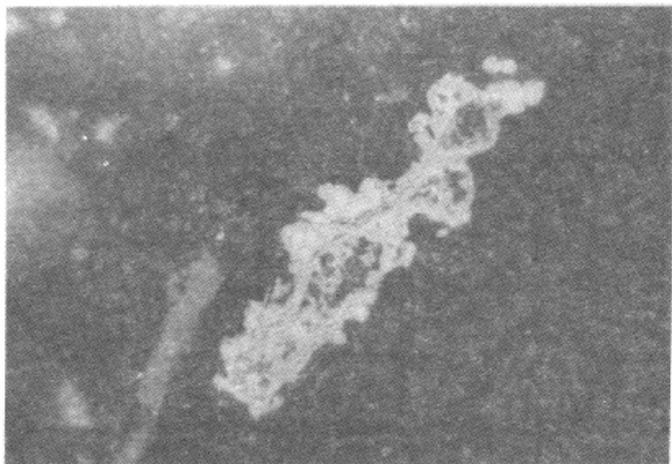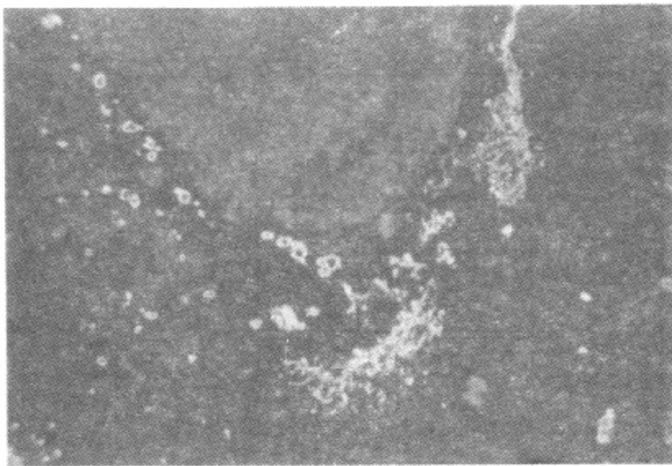
Figure 4a. Chalcopyrite formed by sulphidation of iron oxide encrusting an abandoned worm burrow. Oil immersion. Field: 3mm.
b. Crustiform orange pink bornite overgrowth on early chalcopyrite. Oil immersion. Field 60 μ.

△Cu percentage is only 8%. Thus there is a significant trend from predominantly chalcopyrite copper mineralization at the mouth of the estuary to predominantly bornite mineralization at its head. However, in the middle reaches there is a mixed zone of bornite-chalcopyrite-organically associated and adsorbed copper.

In order to further investigate these indications a number of polished sections of representative sediment types were examined. The major problems encountered during the study of the sulphide mineralogy of Restronguet Creek sediment included the submicroscopic size of most monosulphide grains, their rapid air oxidation characteristics and the soft, poorly crystalline nature of many polysulphides. However careful resin impregnation of undried sediment under vacuum has yielded sufficient polished sections for limited microscopic and S.E.M. study. These techniques are not yet sufficiently developed for preservation and subsequent identification of iron monosulphide species.

On the broadest scale the mineralogical trends indicated by the geochemical studies discussed above are confirmed but there are many local departures where pyrite, chalcopyrite and bornite exhibit relationships which are almost totally dependent on local sediment permeability and metal supply parameters. In intertidal flats sediment bornite is, as expected, the dominant copper sulphide but in many samples it is merely overgrowing and replacing early formed chalcopyrite grains giving fine encrusting, garland and atoll textures.

In one example, (Fig. 4a), chalcopyrite is very prominent and probably initially formed by partial dissolution and sulphidation of cupriferous iron oxide encrusting the walls of an abandoned worm tube. It is now being overgrown and replaced by crustiform orange pink bornite which is clearly visible in the enlargement, Figure 4b. Similarly a woody organic fragment encrusted first by chalcpyrite which is being overgrown by bornite is illustrated in Figures 5a-d. In this series (a) is an S.E.M. electron image of the grain, and (b) and (c) are semi-quantitative energy dispersive Fe and Cu assays. The tracing (d), demonstrates that the bornite occurs as peripheral encrustations and replacement grains partially enclosing chalcopyrite.

Elsewhere in upper Restronguet Creek and especially in oxidized saltmarsh sediment there is evidence that simple supergene replacement of detrital chalcopyrite by purple bornite and bluish chalcocite is the dominant process. These minerals tend not to form crustiform or atoll-type textures. Replacement of pyrite framboids by copper minerals was not seen and there is no convincing evidence that encrusting, poorly crystallized pyrite is replaced either. A few grains and masses of pyrite which might arguably have been detrital and which were encrusted by iron oxide are further overgrown by chalcopyrite or bornite but this occurrence could have been due to movement of these particular grains from an oxidizing to a more cupriferous reducing environment.

## Conclusions

These studies have identified Restronguet Creek as an important example of the syndiagenetic formation of a zoned assemblage of iron and copper sulphide and iron silicate minerals. Assay values of 6% Fe and 0.4% Cu are obtained in some sedimentary sub-environments and these are between two and ten times the levels recorded in similar but artificially unpolluted sediment from the nearby Tresillian River. Some values are two orders of magnitude greater than equivalent sediments from the E. Anglian Wash.

Garland and atoll-type textures clearly indicate the early diagenetic development of copper sulphides whose precise mineralogy is controlled by supergene processes in the upper estuary and by local metal supply and sediment permeability in downstream areas.

Replacement of pyrite by copper sulphide is not an important process at least during the early stages of diagenesis.
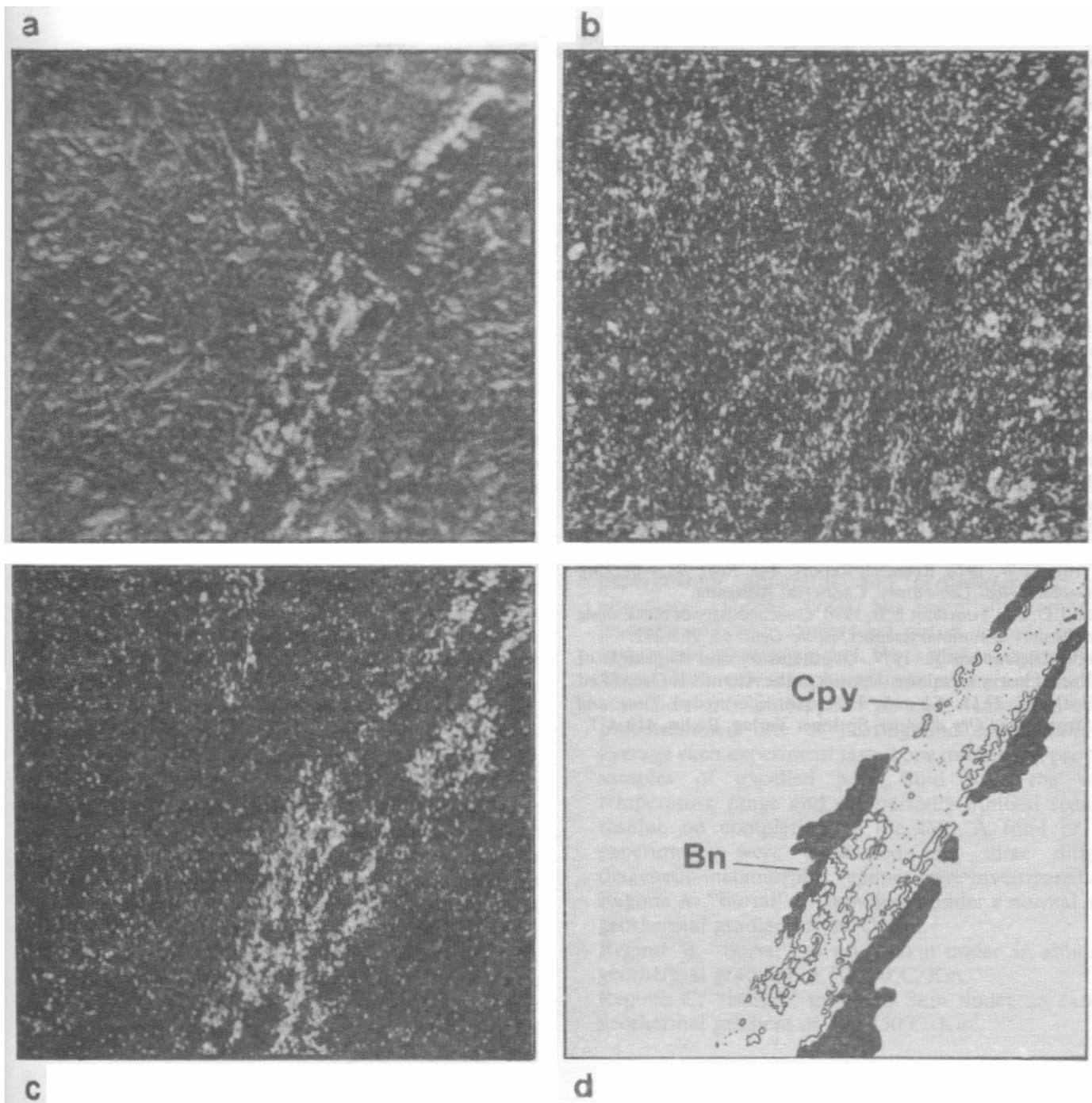
Figure 5a. S.E.M. electron image of copper sulphides encrusting
a woody organic fragment. Field 250 μ.
b. Iron assay of a.
c. Copper assay of a.
d. Interpretative tracing of a. Bn = bornite, Cpy = chalcopyrite.

As far as the author is aware this is the first published report of such an areally extensive occurrence of significant concentrations of diagenetic copper sulphide minerals in Recent sediments.

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# Studies on Fal Estuary sediments II: Artificial diagenesis experiments

M.G. THORNE

Thorne, M.G., 1983. Studies on Fal Estuary Sediments: II, Artificial diagenesis experiments. *Proc. Ussher Soc., 5,* 445-447.

Samples of Restronguet Creek metalliferous sediments have been compressed and heated to simulate diagenesis and incipient burial metamorphism at "depths" of up to 3Km and temperatures of up to 200°C. Under normal geothermal gradients of 30°C/Km expelled pore fluids contain an average of l12ppb dissolved Cu and 320ppb Zn which is close to a range of published mean values for non-mineralizing oilfield waters. Under abnormal geothermal gradients which would produce temperatures between 90°C and 200°C at 3Km the pore fluids contain between one and five ppm dissolved copper and Zinc. These concentrations are similar to those in mineralizing geothermal brines from the Slaton Sea, U.S.A., Cheeked, USSR and the Red Sea.

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## Introduction

As discussed in the introduction to Part I of this report, (Thorne, 1983) certain ancient metal sulphide ore deposits are generally considered to be syndiagenetic with the enclosing sediments. However hypotheses first published by Davidson (1962) Levering (1963) and Noble (1963) suggest that mineralization may not be completed in the diagenetic environment and that the grade of any such mineralization is likely to be enhanced by deposition of metal sulphides from ascending formation fluids either expelled from "source beds" within the sedimentary sequence itself or derived from exogenous sources. Numerous applications of these hypotheses have been made to the genesis of a variety of ore deposits some of which are reviewed by Brown (1981). The literature on theoretical and small scale laboratory studies of these processes is voluminous with reports by White (1971) and Bubela *et al* (1975) being typical of the theoretical and laboratory approaches respectively, However, the author is aware of only one study, by Mossman and Heffernan (!978), in which natural unconsolidated metalliferous sediment has been heated and compressed to study its response to late diagenesis and the onset of burial metamorphism. These authors do not mention the nature of porefluids expelled from this compacting sediment which was from an hydrothermally active Red sea deep basin. If the contribution of locally generated mineralised pore fluid to the formation of sulphide ore deposits in siliciclastic rocks is to be evaluated then potential source rocks must be heated and compressed in order to demonstrate whether they will yield metal rich pore fluids and if so at what temperature and pressure and with what induced mineralogical change in the partially lithified residue. In this study metalliferous Restronguet Creek sediment, (Thorne 1983) is considered as a source rock and has been experimentally heated and compressed in order to investigate these possibilities.

## Experiments simulating diagenesis and metamorphism

Samples of metalliferous Restronguet Creek sediments have been compressed and heated to simulate diagenesis and incipient burial metamorphism at "depths" of up to 3Km and temperatures of up to 200°C. Details of the experimental apparatus and procedures are given elsewhere, (Thorne 1983), but essentially comprise a pressure cell with compacting piston, sample drainage system and a heating jacket. Wet sediment cores of up to 10cms in diameter are placed in the pressure cell and after expulsion of entrapped air are subjected to a predetermined rate of heating and compaction. On average each experiment lasted one month and provided samples of expelled pore fluid over the whole temperature range and the partially lithified sediment residue on completion of the test. A total of nine experiments were conducted and three different diagenetic-metamorphic regimes were investigated:-

Regime A: "burial" to 3km depth under a normal mean geothermal gradient of 30°C/Km.

Regime B: "burial" to up to 3km under an abnormal geothermal gradient of 50-100°C/Km.

Regime C: "burial" to up to 3km under an extreme geothermal gradient of 100-150°C/Km.

## Experimental results

Figure 1 illustrates the behaviour of copper and zinc concentrations in expelled pore water as an average of all nine experiments over the range 20-90°C. Copper and zinc are both initially depleted from pore waters during diagenesis but with rising temperatures these elements are progressively released from sediments. Typical mean pore water concentrations attained during this phase are l12ppb Cu and 320ppb Zn.
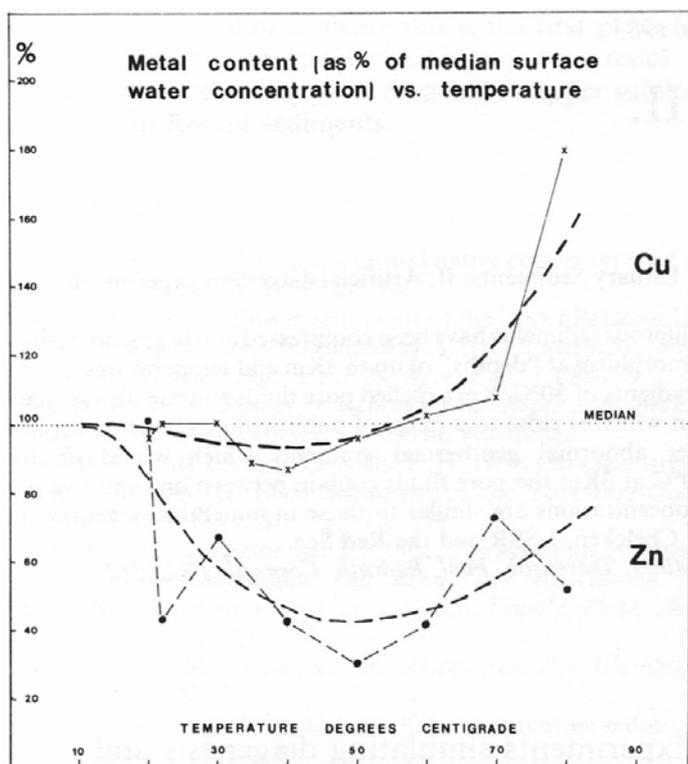
Figure 1. Variation in copper and zinc concentrations in expelled pore water during simulated diagenesis experiments

samples which were heated to lower temperatures both chalcopyrite and bornite have been induced to recrystallize from submicroscopic grains into well developed encrustations, rims, garlands, botryoidal masses and stellate aggregates without alteration or replacement. There was no evidence at all of replacement of any variety of pyrite by a copper sulphide mineral.



Figure 2. Garlands of chalcopyrite and bornite formed in response to high temperature compaction

After extreme thermal treatment of two samples at temperatures of up to 200°C mean zinc and copper concentrations of between 1 and 5 parts per million were obtained. The highest concentration obtained in a single test was 60ppm Zn at 200°C. High concentrations of chlorine (20,000ppm) and boron (tens ppm) were also recorded.

Examination of polished sections of partially lithified sediment indicates that the sulphide mineralogy of most sediments has changed considerably in response to the imposed experimental conditions. Even under normal geothermal gradients and at temperatures of less than 90°C recrystallization of iron monosulphide to pinkish brown poorly crystalline pyrite was swift. Early diagenetic pyrite framboids were transformed into enlarged atoll textured pyrite grains with disordered internal structure and 5-10µthick external encrustations. These grains were resistant to air oxidation but were decomposed by dilute HCl.

In many samples intricate garlands of sulphide minerals developed during high temperature tests; and there are indications that some were associated with vapour bubbles, and others may have been the product of advancing diffusion fronts. (Fig. 2). Chalcopyrite and bornite were the most common sulphides in these assemblages and in some chalcopyrite was replacing early formed bornite. This replacement reaction seems to have been promoted by high temperatures and was most advanced in samples heated to above 120°C. In other

## Discussion of results

These results demonstrate that copper and zinc enriched Restronguet Creek sediment will develop macroscopic copper sulphides during burial diagenesis and metamorphism. However zinc sulphides were not positively identified in the partially lithified sample residues. The evidence indicates that in the majority of the nine tested samples early formed bornite is converted to chalcopyrite either by replacement or recrystallization following dissolution. Certainly there is more than sufficient iron available in the tested sediments for this reaction to proceed to completion given appropriate thermodynamic conditions. This evidence initially appears somewhat at odds with findings from many ancient syndiagenetic copper deposits including White Pine, Michigan, USA and on the Zambian copper belt, where chalocite and bornite are very important if not the dominant copper minerals. However, it is possible that for some parts of some Zambian deposits at least the predominant late diagenetic mineral in most silt grade carbonaceous sediments was chalcopyrite and this was subsequently partially replaced by bornite and chalcocite on the margins of the deposit, Van Eden (1974).

Concentrations of hundreds of parts per billion copper and zinc dissolved in expelled porewaters are quite unremarkable when considered in the light of analytical results from formation fluids extracted from various N. American oilfields, (Rittenhouse et al 1969), and from the W. Canadian sedimentary basin (as summarized in

Hitchon, 1977). These naturally occurring formation fluids give mean dissolved metal concentrations of 120ppb for Cu and 300ppb for zinc. Thus under normal geothermal gradients metalliferous Restronguet Creek sediment would be unlikely to yield potentially mineralizing pore fluid at burial depths of less than 3Km. However at temperatures between 90°C and 200°C the experimental results show that expelled pore fluids develop dissolved Cu and Zn concentrations similar to those quoted by Hitchon (1977) for the Slaton Sea, U.S.A., Cheeked, USSR and Red Sea geothermal brines but at only about 20% of their reported chloride ion concentrations. These brines are producing metal sulphide deposits in the enclosing country rocks. Therefore it is reasonable to assume that the expelled high temperature Restronguet Creek pore fluids would also be capable of forming sulphides given a supply of reduced sulphur and a suitable depositional environment. Although no actual measurements were made in the present study it should, theoretically, be possible for these expelled solutions to transport a portion of the required fixative sulphide along with the dissolved metal, Rose (1976) and Nriagu and Anderson (1970). This will be investigated in continuing studies.

## Conclusions

This report demonstrates that simulated burial to 3Km of mineralized Restronguet Creek sediment under a normal geothermal gradient produces copper and zinc concentrations in the expelled pore fluids which are comparable with results from non-mineralizing oilfield formation fluids. However, with an abnormally high geothermal gradient or presumably a greater depth of burial potentially mineralizing pore fluids are produced. These fluids are not brines, have no "hydrothermal" component and yet develop dissolved metal concentrations similar to those from the Slaton Sea, USA, Cheeked, USSR, and the Red Sea geothermal deeps.

These studies are the initial stages of continuing research in this field. Work is currently in progress to determine whether different types of mineralized sediment produce different expelled pore fluids and to establish whether other types of pyritic but copper poor sediments can be mineralized by them.

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# The Alderney Sandstone: an alluvial fan

D.G. RATHAN

Rathan, D.G., 1983. The Alderney Sandstone: an alluvial fan. *Proc. Ussher Soc., 5,* 448-453.

The Alderney Sandstone represents a Lower Palaeozoic semi-arid alluvial fan. A typical debris flow-streamflood - braided stream sequence is recognised, with proximal sheetflows, an intersection zone, and reduced discharge on to the pediplain.

Four lithofacies are defined: Facies a are basal beds of debris flows and sheetflood deposits, formed in the piedmont zone; Facies b are streamlain orthoconglomerates at the fan apex; Facies c is where the incised channel reaches the fan surface, giving sheetfloods; and Facies d represents sweeping discharge across the distal reaches, and constitutes the major part of the total sequence. These deposits are certainly post-Cadomian; the best estimate of their age is Lower to Middle Cambrian. *D. G. Rathan, Geology Division, Department of Environmental Sciences, Plymouth Polytechnic, Drake Circus, Plymouth PL4 8AA.*

## Introduction

The Alderney sandstone is a succession of continental arkosic and subarkosic sandstones, conglomerates and coarse siltstones, approximately 800 metres thick; it lies upon a Precambrian crystalline basement. A similar sequence occurs on the nearby French mainland. A Cambrian age was suggested by Hill (1889), and has been taken up by Plymen in 1922, and by Pudsey (1977). An inferred Ordovician cover has been completely removed. The beds are exposed at the Western Outlier in the south-east of the island; at the main outcrop, in the north - east; and on Burhou, a small island some four kilometres to the north-west of Alderney. A further outcrop (Hill, 1889) occurs at the Casquets, another eight kilometres to the west. On Alderney itself, the Western Outlier comprises a short succession above the Western Granodiorite; the main outcrop lies above the Central Diorite. Inland exposure is mostly in the quarries; coastal exposure is excellent, although the south-facing shoreline is of cliffs, and generally very difficult of access. Generally, the beds dip to the east-south-east, at an average angle of 30°; deformation is limited to gentle monoclines, a thrust and considerable faulting. No evidence of metamorphism was observed.
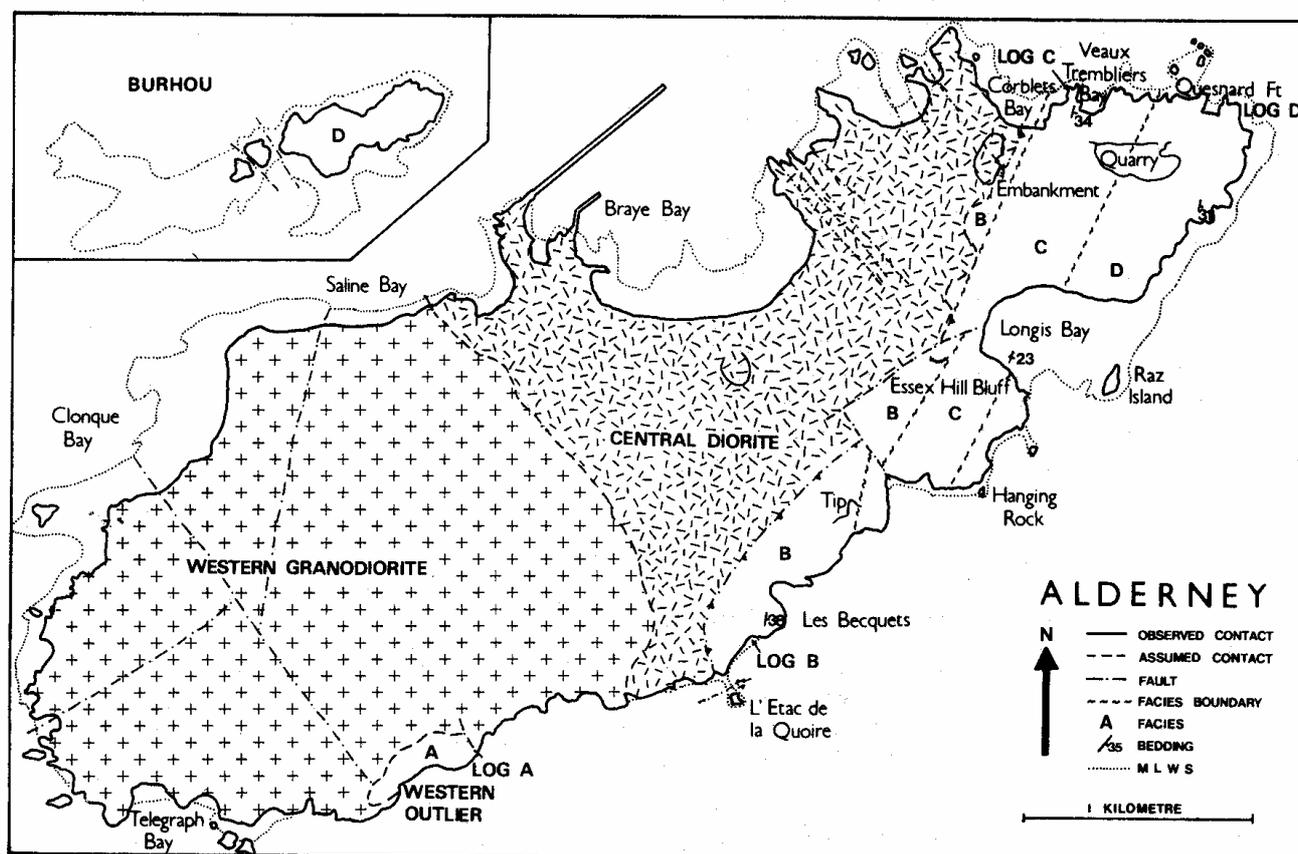


Figure 1. The general geology of Alderney, and localities mentioned. (Note, the exact geographical limitation of facies cannot be defined). A - Paraconglomerate and parallel-laminated sandstone facies, B - Orthoconglomerate facies, C - Sheet sandstone facies, D - Crossbedded facies.

The Alderney Sandstone has been split into four distinct lithofacies (Fig. 1). These are Facies a, the paraconglomerate and parallel-laminated facies, Facies b, the orthoconglomerate facies, Facies c, predominantly sheet sandstone facies, and Facies d, the crossbedded facies (Fig. 2).

Facies a is approximately 15m thick. It commences with some 3m of breccias, above an observed sedimentary contact; these are overlain by the dominant sheetflood sandstones.

Facies b begins above a thin succession of quartzitic sandstones, at the stratigraphically lowest part of the main outcrop, above a thrust contact. This facies is typically seen at Les Becquets and the Essex Hill Bluff; it consists of streamlain feldspathic sandstones and conglomerates.

Facies c occurs, broadly, from Corblets Bay to the Railway Quarry, in the north-east, and along the southern margin of Longis Bay, at the southern end of the strike. In this latter region, evidence for the presence of this Zone is not entirely clear. The lithofacies is identified by a predominance of sheetflood sandstones.

Facies d extends from the limits of Facies c to the end of the island, and also forms Burhou. Typically, these units are relatively mature subarkosic sandstones, in bedform and flood-plain horizons. At the top of the succession, the outcrops of Fort Quesnard and Fort Houmet Herbé, and on Burhou, are classified as a subfacies, since there is distinctive sedimentological variation from the lower beds.

## Facies a, the paraconglomerate and parallel-laminated facies

Facies a outcrops at the Western Outlier. It is the only one of the four facies which lies with a normal sedimentary contact upon the basement; Facies b, c and d have an overthrust relationship. The paraconglomerates marking the base are the most immature sediment seen in the Alderney sandstone; because of this, together with the above mentioned field relationships, it is considered the basal unit of the Alderney sandstone.

The granodioritic and felsitic basement is weathered and exfoliated; it is overlain by an accumulation three and a half metres thick, of matrix-supported lithic breccias. The Clasts, mostly fragments of felsite, but also of quartz and granodiorite are subangular but frequently display high sphericity. They tend to decrease in size upwards and are contained in a poorly-sorted coarse, angular and compositionally-immature matrix. There is a conspicuous absence of cross- stratification, scouring, imbrication and planar-stratification.

Above these comes a texturally distinct type of beds. Their principal characteristic is a fining-up fabric, from a pebbly base to a medium-grained sandstone. Planar cross-bedding and frequent cross-laminations occur at the top of each of these units, with a planar laminated

fabric prevailing throughout most of each unit. The extensive lateral continuity of these planar-bounded units highlights the absence of channelling or scouring. Occasionally, thin red micaceous siltstones are present.

A stream-transportation mode is precluded by the poorly-sorted nature and large clast size of the basal matrix-supported breccias, combined with the. complete absence of sedimentary bedform structures. As a result, the interpretation must be one of high-viscosity debris-flows.

The fining-up nature of the planar-bounded units indicates a fluctuating flow regime, with regular flood events and subsequent waning flows. The absence of channelling strengthens the conclusion that these deposits were formed by sheetflood processes.

## Facies b, the orthoconglomerate facies

On the main outcrop, the sequence begins above a thrusted base, which lies upon the Central Diorite. The thrust is interpreted as one of small movement, along the original sedimentary contact. Evidence for this concept lies in the large rounded boulder-sized clasts of Central Diorite composition which are contained within a finer fault-gouge material above the thrust plane. An acute angle exists between the bedding and the thrust plane.

The succession commences with a thin sequence of quartzitic sandstones, which display fining-up fabrics, planar cross-stratification and, planar laminations. This is followed by a thick continuous sequence of partially-stratified orthoconglomerates, consisting of rounded pebble-sized clasts held in a gritty matrix. Large-scale trough cross-bedding is present, seen on cross-cutting scour surfaces. There is a high mean feldspathic content, and feldspar clasts range up to 7cm across; they are occasionally euhedral, but more commonly are moderately well-rounded. Up the succession, the rounding of the clasts and a general decrease in clast size allow an arbitrary demarcation of the top of these beds. The identification of this coarse facies as orthoconglomerate is based upon its stratification, following Bluck's classification (1967). He stated that such beds may be deposited as the result of continuous or multiple streamflows incising the proximal facies of a fan, and such an interpretation is favoured here. These coarse clastic deposits, with their poor sorting and yet, in general, well-rounded clasts, are inferred to occur as the consequences of frequent reworking by high-velocity flows, such as that of a transient sweeping discharge in a deep channel incision across the relatively short arc of the fan apex. Stream reworking will have acted upon the debris-flow accumulations upstream, rounding the clasts and winnowing out the fines. Variation in rainfall must have been the cause of the localised increases in grain size. The depth of the major channel decreases down the fan (Hooke, 1967), with a related increase in channel width, towards the intersection point, defined where the channel base meets the fan surface and the beginning of the next facies.
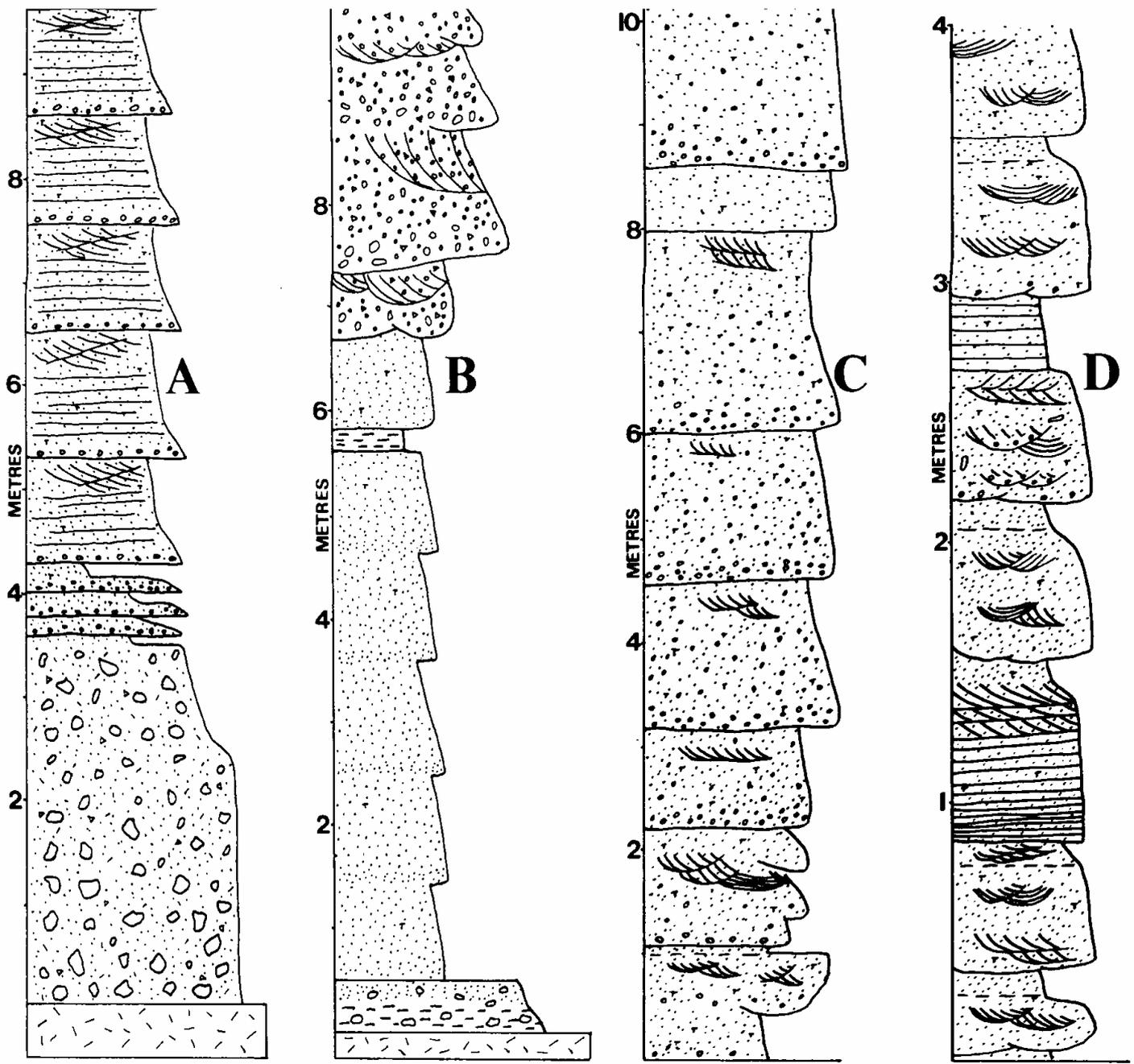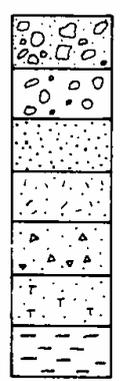
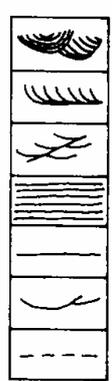Figure 2. Simplified graphic logs of the major lithofacies

## Facies c, the sheet sandstone facies

The next sequence of beds outcrops on both the northern and southern coastlines, and can be examined in detail at Corblets Bay and Veaux Trembliers Bay. The varied nature of the fabric and structure of the sections within this facies renders it distinctive. Most abundant are the fining-up sheet sandstone units, similar to those seen in the Western Outlier succession, although texturally more mature. Thin fining-up (10-40cm) horizons of oxidised siltstone and fine sandstone frequently inter-digitate with these. Also present are units which are recognisable in texture from both the underlying Facies b and the overlying Facies d.

On an alluvial fan, the area upstream of the intersection point displays considerable trough cross-bedding, in stratified gravel horizons; these features are represented in the Alderney deposits within the sections of the basal part of this lithofacies, at Fort Coblets, the Railway Embankment and the Rubbish Tip. Within the upper parts of this sequence occur bedform structures which are typical of the immediately-following Facies d, these are best seen at the Railway Quarry and Hanging Rock. At these locations, however, and particularly between them, fining-up sheet sandstone units predominate; these are sheetflood deposits. A floodplain interpretation follows from the ubiquitous persistence and fining-up nature of the finer-grained horizons.

The recognition of these sheetflood units allows the identification of this facies, and its relationships to the fan surface. Below the intersection point the wash is dispersed evenly across the surface in a sheetflow mode. During the growth of the fan, the intersection point migrates upstream, although it is bound to vary backwards and forwards, Over relatively shorter periods. This variation results in a zone in which facies units typical of the succession above and below this level are found to interdigitate: an Intersection Zone is thus characterised.

## Facies d, the crossbedded facies

The sedimentary succession of the main outcrop is completed by this facies; it comprises the majority of the coastal exposure. The general structure is still that of a regular inter-digitation between coarse-grained streamlain deposits, which are up to a metre thick, and the finer-grained floodplain horizons, which are usually about ten centimetres thick. The streamlain deposits however show a structural diversity. The most abundant mode is that of a scour-and-fill sequence, consisting of large-scale scour structures with laminations parallel to the scour base, trough cross-bedding with steeply dipping foresets, and laterally impersistent silt drapes. Planar crossbedding is also abundant, usually in the form of cosets, which are sometimes confined to their own units, but which are occasionally found with trough crossbedding. Some unusual forms of planar

crossbedding are present: for example, herringbone wedge-shaped planar cross-sets (10-15cm set height), and large, scale single-unit planar cross-stratified sets (l.0-1.5m set/unit height). Channel scours also occur, with cross-stratified sandstone infills. The laterally persistent flood-plain units sometimes showed ripple cross-lamination, but, more frequently, consisted only of a fining-up fabric.

The interpretation of these structures relies, quite considerably, on the work of Miall (1977). The bedform structures of planar and trough crossbedding may be considered as the result of sandwave and scour-and-dune migration within the stream. Silt drapes represent low water stages. Herringbone wedge-shaped planar cross-sets have recently been identified (Bluck, pers. comm.) as records of backflow ripple migration, due to turbulence at the leeward end of a bar, during flood conditions. Large-scale single unit planar cross-stratified sets can be readily identified as the products of downstream bar migration.

These deposits are envisaged as having been laid down under the control of a sweeping braided discharge on the distal parts of a semi-arid fan. It is considered that a tentative correlation may exist between the separation of the flood-plain horizons and the period of sweep of the discharge. A considerable braided system must have been established, to give these impressive migratory bar structures. A distal position on the fan is a logical consequence of the nature of the succession; this is supported by the textural and compositional maturity trends.

*A Subfacies of d.*
Two further locations provide additional information on the nature of the distal reaches of the fan; the first is at the highest part of the braided stream section, and the second is on Burhou. At the highest part of the braided stream section on Alderney, a few subtle distinctions exist in lithology, in unit types, and in structures. As expected, there is increased compositional and textural maturity, still with the interdigitation between bedform and floodplain horizons. Bar migratory structures are apparent. Thin vertically-accreted floodplain deposits are sometimes missing; their place is taken, preferentially, by thicker horizontally-stratified units which are interpreted as sheetflows during exceptional flooding. The most conspicuous departure from the norm, however, lies in the more abundant and larger channel scours. This fits well into the larger picture: the larger and more persistent channels commonly found further downstream would have the greater preservation potential.
Burhou probably displays the highest section in the sequence, inferred from maturity criteria and (Pudsey, 1977) structural grounds. The main features are those of a significant, and rather unexpected, decrease in preserved channels. Migratory bar structures are still present, within a braided stream lithological fabric. An increase in the number of sheetflood units is apparent.

It could well be that a reduced volume of discharge was responsible for the decrease in channels, as the result of internal drainage and evaporation: Another interpretation is favoured, however, since sheetflood deposits are present, and these suggest occasional periods of considerable outflow. The change in slope, from fan base to pediplain, would give a general reduction in flow velocities; the sheetflood episodes may owe much of their nature to the relative change in confinement which results from the much-reduced dominant slope direction of a pediplain.
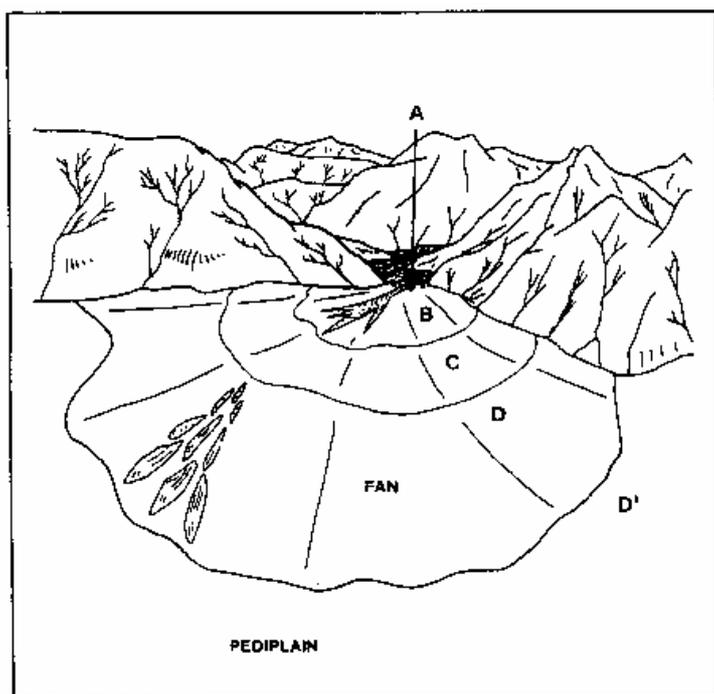


Figure 3. A reconstruction of the depositional environment of the Alderney Sandstone.
A - Debris flows and localised sheetflows, B - Multiple or continuous, incised, streamflows, C - Sheetflow dominated intersection zone, D- Fan sweeping braided streams, D' - Ephemeral pediplain braided-discharges.

## Model

The recognition of the various facies of the Alderney Sandstone allows a dynamic reconstruction. Erosion of the Cadomian mountain belt has already been envisaged by, *inter alia,* Sutton and Watson (1970), as being the source responsible for the development of the Alderney Sandstone.

An alluvial fan is suggested here, as the environment of deposition. The basic textural and structural evolution which would develop within the sequence of a growing fan is debris flow to streamflood to braided stream (Hooke, 1967); these facies, and the sequence, have been established on Alderney.

In summary, the initiation of this sedimentary sequence is marked by the development of debris flows, from talus heaps within the mountain range. Storms caused flash floods, giving sheetflows in the unconfined regions of the piedmont zone; these are preserved as sheetflood deposits overlying the debris flow deposits. The reworking of these deposits, especially the debris flow accumulations, led to the gradual encroachment of the stream-lain conglomerates. Their deposition initiated the fan; as it grew, the various zones became distinguishable (Fig. 3), as the sheetflood-dominated intersection zone, the fan-sweeping braided stream zone, and the ephemeral pediplain braided-discharge zone. Effective mountain belt retreat, by the process of surface downwasting, brought each facies to overlie its predecessor in a transgressive manner.

## Discussion

Doré, in 1972, showed Alderney as part of an emergent Cambrian landmass, the result of uplift during the Cadomian orogeny. He suggested transgressive shorelines for the Cambrian seas, forming embayments running westwards into the Cadomian uplands. On Alderney, the palaeocurrent flow directions had a considerable variance, but did show a southerly flow, possibly towards an embayment. Transgression during Lower Ordovician times on to the lowlying regions would result in the alluvial fan facies grading upwards and laterally into a marine facies; this picture is in general agreement with the presence of Ordovician slates (Fily, 1972) on the sea-bed to the south-east. Marine incursions are absent from the Alderney succession, so that a palaeogeographical reconstruction points to a Lower or Middle, rather than Upper, Cambrian age for the Alderney Sandstone.

The proximity of Alderney to Britain, and the basic trend (ENE-WSW) of the Cadomian orogeny, suggest that, in southern Britain, beneath the thick Mesozoic and Cenozoic deposits, any remaining Cambrian strata would be of the same semi-arid continental style, overlying a crystalline basement.

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# The evolution of a barrier-lagoon system -a case study from Start Bay

## C.R. MOREY

Morey, C.R. 1983. The evolution of barrier-lagoon system - a case study from Start Bay. *Proc. Ussher Soc. 5,* 454-459.

The shore of Start Bay is formed by an intermittent fringing gravel barrier 9km in length, This rests against a degraded cliffline of pre-Holocene age. One section of the barrier - Slapton Sands -represents the major accumulation of gravel and this occupies the site of a former estuary. Sections of the barrier have dammed the lower reaches of valleys that drain to the coast to form lagoons or wetlands that provide the stable hydrological conditions suitable for the formation of fen peat. Investigation of the back barrier sediments at Slapton indicates that they are predominantly of fresh water origin and shows that major beach movements shortly before 3,000 years B.P. led to the closure of the lagoon. Subsequent overwash during the first four centuries A.D. has raised the height of the barrier crest and initiated a second cycle of sedimentation. A feature of the present situation is that lagoons and coastal marshes are "perched" above mean sea level and are imposing a progressive rise in the base levels of local catchments.

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## Introduction

Start Bay lies on the eastern shore of the South Hams peninsula in South Devon, England (Fig 1): The coast- -line is dominated by a fringing barrier composed of granules and small pebbles with little interstitial sand. This feature is approximately 9km long and links a series of nodal headlands which behave as natural groynes controlling its form. For much of its length the barrier abuts a cliffline which is probably polycyclic but which was blocked out by the end of the Ipswichian interglacial (Morey 1976). A series of six pre-Holocene valleys have been dammed to produce back barrier lagoons. These lagoons contain either open (fresh) water or reedswamp dominated by *Phragmites communis. The* offshore zone has been investigated by the Institute of Oceanographic Sciences (Hails, 1975 a,b) and the neighbouring area to the east by Clarke (1970). The back barrier sediments have been examined by Morey (1976, 1980). Taken together, these works appear to offer a detailed record of coastal evolution in the area since approximately 9,500 B.P.

Physiographically the rock floor of Start Bay is an extension of the landscape of South Devon submerged during the Holocene transgression. It is overlain by a veneer of recent marine sediment of variable thickness. Bed rock outcrops on former interfluves while valleys have up to 28m of Holocene sediments. Palaeoenvironmental work by Lees (1975) has shown that these deposits may be divided into two sections. The lower bay deposits overlie bed rock and consist of units representing estuarine, salt marsh and hyposaline conditions apparently associated with a transgressive coastline. In the vibrocores studied, a thin layer of modern offshore
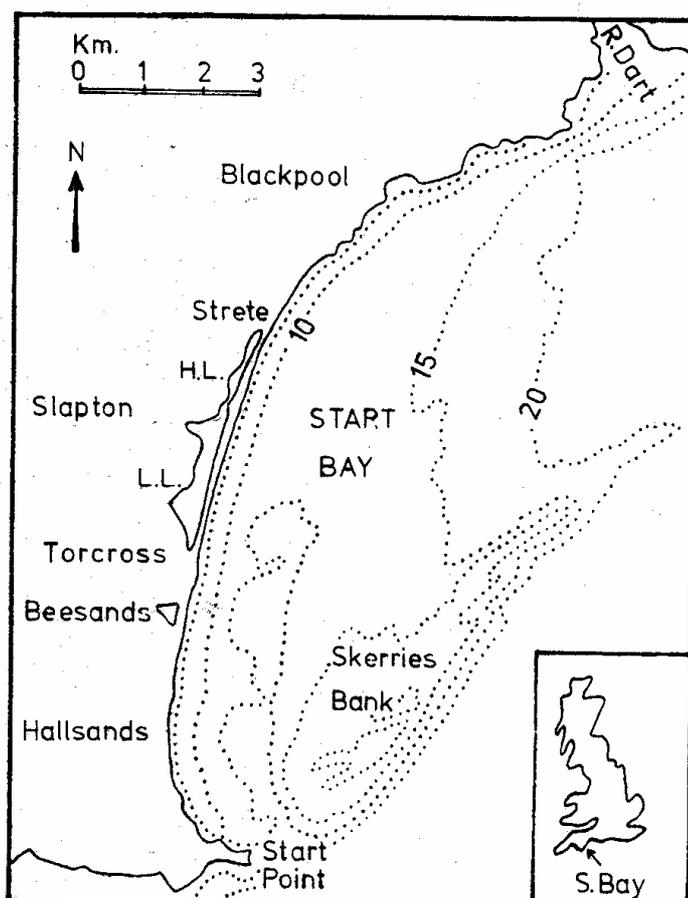


Figure 1. The study area showing the location of the main sites referred to in the text. H.L. = Higher Ley, L.L. = Lower Ley. Bathymetry in metres below chart datum.

sands overlies the lower bay deposits and has been referred to as the upper bay deposits by Hails (1975a). Although fine grained facies are preserved throughout the area relict barrier deposits are rare and only one site with fresh water peat has been reported - this from a site on the inner margin of the Skerries Bank about 2km east of the present shoreline at Hallsands. The absolute date for this peat (8,018 $\pm$ 60 B.P.) is highly significant since it appears to have developed under fresh water conditions, probably in a back barrier lagoon, and it therefore gives an upper limit for sea level at the time that the Holocene transgression began to spread into Start Bay.

Apart from the site on the Skerries Bank all of the occurrences of relict barrier deposits recorded by Hails *(op. cit.)* are situated close to the present coastline. A series of vibrocores along a line from Strete to Blackpool encountered relict gravels, while fen peat 40cm thick dated between 4,767 $\pm$ 45 B.P. and 4,302 $\pm$ 45 B.P. was recorded from a borehole at Beesands sited near the crest of the existing barrier. It therefore appears that a substantial gravel structure, or a series of structures, existed slightly seaward of the present coastline at around 5,000 B.P. and was stable enough to allow peat to accumulate behind it for several hundred years.

An additional factor to be considered is that, although an area of moderate wave energy, Start Bay falls within the macrotidal category in the classification of Davies (1964) and the suggested modification by Hayes (1979). In such a tide dominated situation barrier formation would be restricted and the history of proto-barriers is likely to have been a repeated cycle of formation, overwashing and destruction as they traversed a landscape of relatively low relief. One would expect lagoons to have been equally impermanent and the apparent absence of relict peats from the offshore zone (Hails 1975a) is readily explained by the fact that the fresh water conditions suitable for fen peat formation require a period of stability behind an established barrier. The failure to discover peat or relict barrier gravels may be due to their destruction during transgression but their absence may also reflect major differences in the depositional environment between the early and late Holocene coastlines.

## The form of the present barrier

Any discussion of the status of the present barrier must take two major factors into account.
(1) The feature is composed of granules and small pebbles of assorted petrological types and there is general agreement among workers from Worth (1904) onwards that Start Bay is a closed system and that 85% of the gravel is of chert or similar material derived from outside the area.
(2) Although preserving an unbroken seaward profile the feature is backed by variable topography so that the continuity of the backslope is broken by a series of headlands that terminate in degraded cliffs. In these sections, which make up almost half of the total length of

the feature, bed rock outcrops and the veneer of barrier gravel rarely exceeds 2m thickness. Between the headlands the linking sections rest on the sediments of former lagoons and boreholes have proved gravel up to
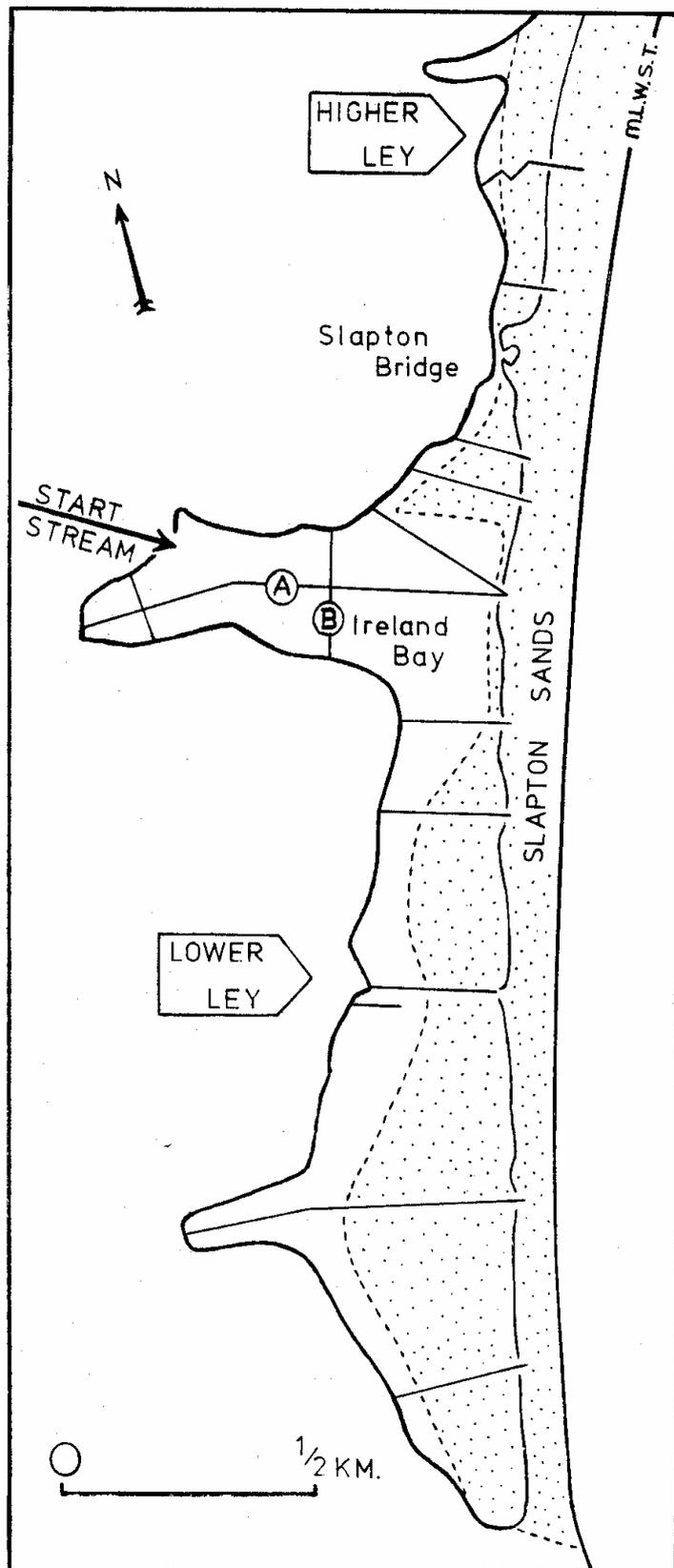
Figure 2. A map of Slapton Lower Ley and a portion of the Higher Ley showing the extent of the washover fans (stippled) and the location of 14 transect lines used in the investigation. Transects (A) and (B) are shown in section in Fig. 3. M.L.W.S.T. = Mean Low Water Mark Spring Tides.

12m in thickness. Only the sections at Slapton Sands and Beesands represent a massive accumulation of gravel and volume estimates have been made by the author (Morey, 1980) indicating that Slapton Sands contain 85% of the material in the Start Bay system shorewards of M.L.W.S.T.

## The back barrier sediments

The coastal lagoons at Slapton and Hallsands have been investigated by the author (Morey, 1980). They provide an interesting contrast in that at Hallsands aggradation of the marshy valley floor has raised the surface of the marsh to the height of the barrier crest while at Slapton an extensive depression is flooded to produce a shallow "Ley", These differences are likely to be related to the sediment supply derived from the surrounding catchments but also to the volume of the lagoon which serves as a sediment trap and to the length of time for which this has been sealed. Slapton Ley appears to have been subject to tidal flushing until about 3,000 B.P. while other valleys on the coast were sealed somewhat earlier.

### Slapton Ley

The area enclosed by Slapton Sands is divided into two sections - the Higher and Lower Leys - by a rock spur at Slapton Bridge (Fig. 2). Investigation of the sub-lacustrine sediments and the adjoining marshlands was carried out using a standard Hiller peat borer. About 180 holes with a maximum depth of 7m were logged along a series of 14 transect lines. The Hiller proved unable to penetrate more than lm into the underlying estuarine muds. The two Leys contain extensive sheets of washover gravel derived from the barrier. The washover sheets of the Higher Ley appear to spread across the whole of the basin and extend south of Slapton Bridge but in the Lower Ley it has proved possible to map the outline of a discrete fan (Fig. 2). The most complete and informative section of the lagoon sediments is located at the seaward side of Ireland Bay where the washover sheets are absent. It is clear that the succession established for Ireland Bay extends to the whole of the lagoon passing beneath the washover sheets to north and south. The location of the washover sheets is significant, supporting the suggestion by Hails (1975b) that the offshore Skerries Bank tends to focus wave energy in the northern and southern sectors of Start Bay leaving the central sector as a low energy zone. In fact the section in Fig. 3 shows that the central sector of Slapton Sands has no washover structures on its inner face and contemporary observation indicates that this sector of the barrier is rarely overtopped by storm waves. Figure 3 shows the major features of the Lower Ley formation which the author has divided into three members which rest upon estuarine muds.

*Estuarine Muds.* These are light grey in colour, predominantly silty, and may well be an extension of the lower bay deposits (Hails 1975a). They pass beneath the barrier where they have been located in boreholes Nos. 82 and 83 by Hails (op. cit.) and extend into the valley of the Start Stream that now supplies the Ley. The muds contain a fauna of low diversity dominated by *Hydrobia ventrosa* with *Ammonia beccarii, Protelphidium anglicum* and *Elphidium* spp. Bivalves including *Macoma* occur on the seaward margin while, as the unit is followed inland, marine species disappear and the muds pass laterally into greyish alluvium. The pattern suggests a salinity gradient with restricted water circulation behind a growing spit or barrier, a situation that would be expected in a small estuary. The tidal entrance has not been located in the course of the investigation due to the extent of subsequent gravel washover sheets but is likely to have been in the southern part of the Lower Ley.

*Basal Silts.* The lowest member of the fresh water facies is a thin brown organic silt with a sharp lower boundary and which passes upwards into the fen peats of the overlying member. The remarkable feature is the
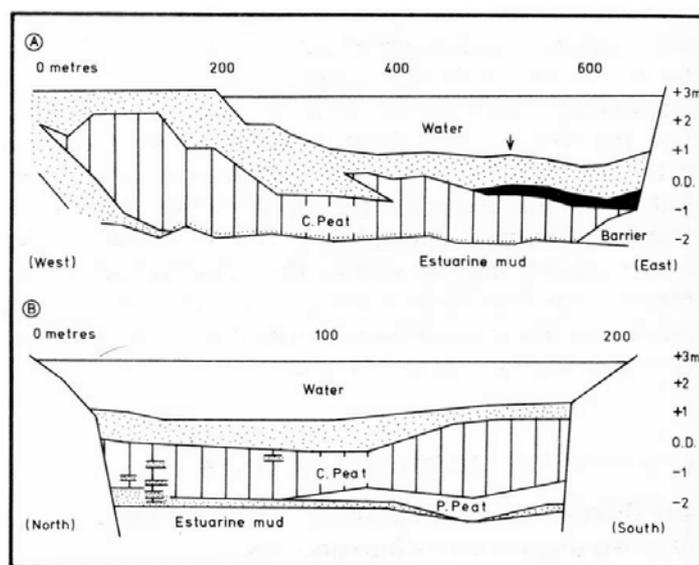


Figure 3. Sections through the sediments of Ireland Bay along transect lines (A) and (B) shown in Fig. 2. Silty units of terrigenous detrital origin are stippled and a sandy washover unit shown in solid black. C. Peat= *Carex* (sedge) peat, P.Peat= *Phragmites* (reedswamp) peat. The arrow on section (A) indicates the site from which the pollen diagram in Fig. 4 was obtained.

sharpness of the lower boundary which suggests an abrupt change in the depositional environment. Pollen analysis has been carried out for the whole thickness of the formation and also shows an abrupt change at this point. The estuarine muds contain little pollen while the basal silts contain an assemblage characteristic of reedswamp *(Phragmites)* containing pools or channels of open water. The pollen of *Chenopodium* (usually regarded as an indicator of salt marsh conditions) is also more important at this stage than any other suggesting that vestigial salt marsh may have existed at this time. In this context the presence *of Phragmites* is interesting- this is a tolerant and adaptable species that colonises a range of damp to aquatic habitats and is uniquely capable of spreading into brackish water to form a transitional community between salt marsh and fresh water swamp. A simplified pollen diagram is shown in Fig. 4.

*Fen Peats.* From the early reedswamp stage there is a steady succession to a *Carex* dominated fen community. The peat is approximately 1.3m thick and ^{14}C dates are available for the base and top (2,889 ± 50 B.P. and 1,813 ± 40 B.P. respectively) so that peat forming conditions

persisted without a break for about 1000 years. The absolute date for the base of the peat makes it possible to infer an approximate date for the closure of the barrier as it is hard to offer any other conditions.

*Lacustrine muds.* Peat formation was terminated by a major marine incursion which spread a sheet of muddy sand across the fen surface. There is some indication towards the southern end of the Ley that this was associated with a gravel fan but the evidence is obscured by later washover. Apart from a few pollen grains and shell fragments the unit is barren but the argument for a marine origin is supported by the fact that it is the only unit to thicken seaward, filling depressions in the peat surface. The event is clearly associated with a major environmental change as it is succeeded by lacustrine muds of terrigenous detrital origin deposited in a water depth of more than two metres. This change is clearly reflected in the pollen diagram with recovery by *Phragmites* and open water species (Morey 1980) and in the diatom flora (Round 1967). The sequence of events may be interpreted as a marine incursion followed by rebuilding of the barrier to a greater height. The major



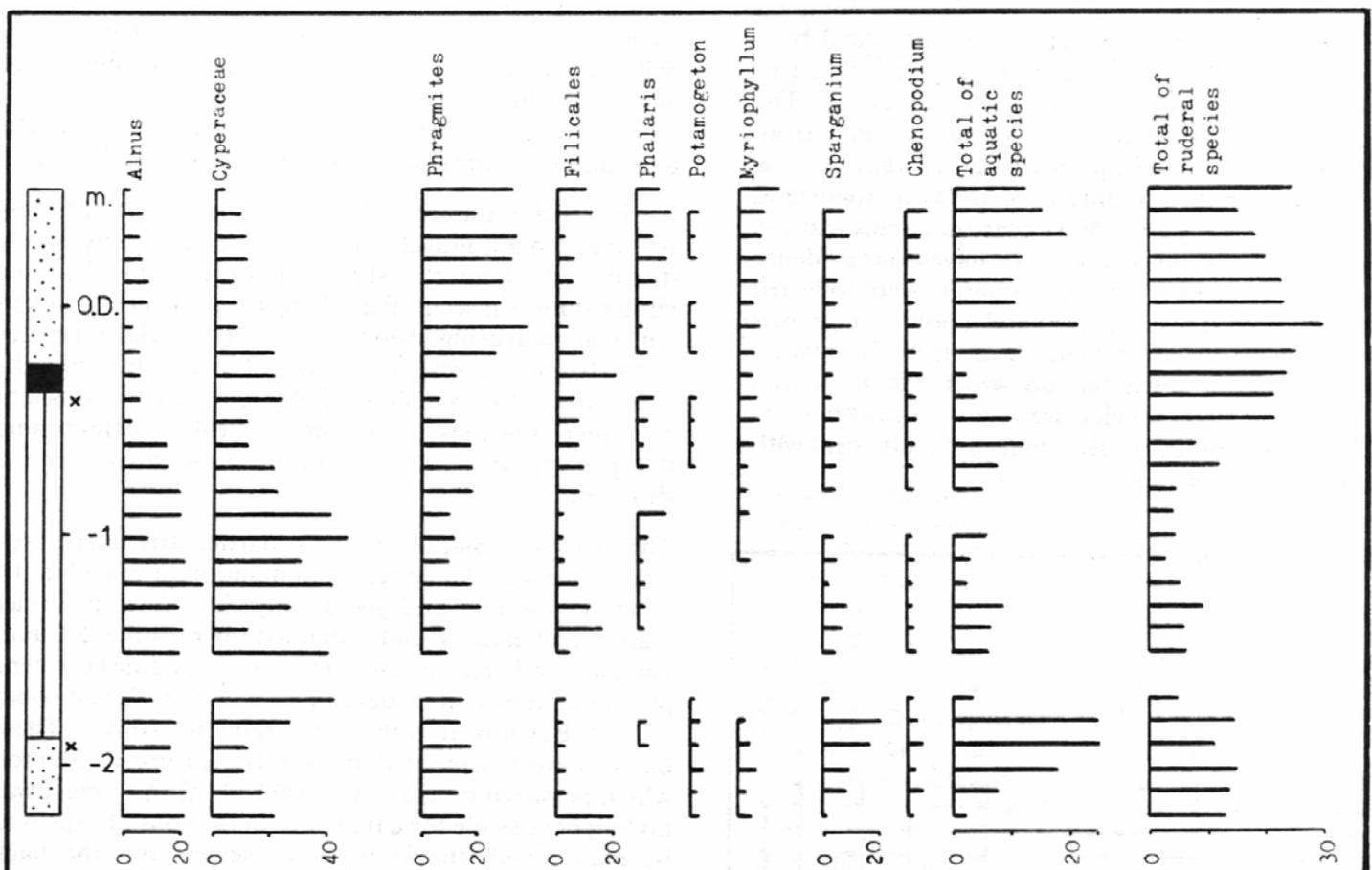Figure 4. Pollen diagram for a site in the centre of Ireland Bay under a 2m water column. The symbols for the sediment column on the left hand side are the same as for Fig. 3 and the positions of ^{14}C dated samples are marked (x). Pollen frequencies along the horizontal axis are calculated on the basis: Total Pollen = 100%. The gap in the pollen record at -1.6m O.D. is due to the presence of wood in the section.

washover sheets referred to above and marked in Fig. 2 interdigitate with the lacustrine sediments suggesting that they were formed since 1,800 B.P., possibly during medieval times.

It will be seen from Figure 3 that the present lake bed is above mean sea level and may be properly described as a perched lake. Increased erosion rates in historical time have caused a steady increase in the terrigenous detrital content of marginal peat deposits and have allowed aggradation of the flat floors of valleys draining to the Ley. It is apparent that the whole Ley system of lake and marginal facies is moving landward and upward through space and time and controlling water tables, sedimentation, and land use in the adjacent valley system.

*Hallsands*

A second site with extensive peat deposits is situated at North Hallsands. This was investigated in the spring of 1974 after gales had stripped the beach of its normal cover and exposed sections showing peat interdigitated with gravel. The lowest organic unit comprised 30cm of dry, laminated, well humified peat, resting on a palaeosol developed from silty aluvium. The upper surface of the peat bore the abraded stumps of trees up to 30cm in diameter and a 14C date of 1683 ± 40 B.P. was obtained from one of them. The tree layer was succeeded by a strongly indurated gravel and it seems likely that tree growth was terminated by a washover event. The absolute date is sufficiently close to that obtained from the uppermost peat at Slapton to suggest that there is probably no significant difference between them and there is the tempting possibility that they represent the same event. Even if this is not so we have clear evidence for beach movements at this time. Samples were collected from the exposed face of the peat and a pollen diagram is shown in Fig. 5 indicating a succession from *Phragmites* marsh to a drier *Carex* fen on which the woodland developed. The tree species were not, unfortunately, identified. The subsequent sediments are a silty peat with *Phragmites* and *Salix* spp.
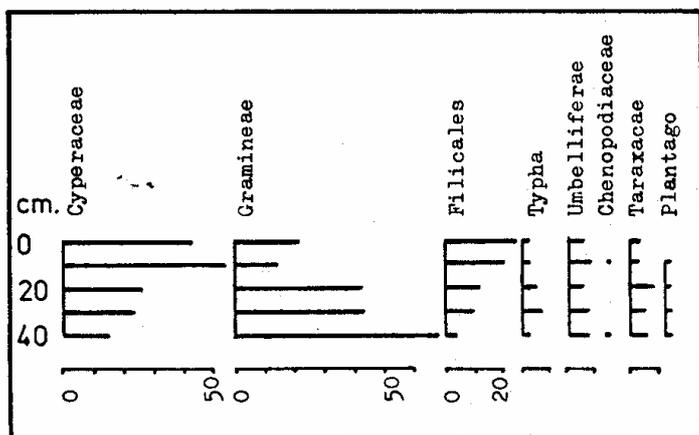


Figure 5. Pollen diagram for a site on the beach at North Hallsands. Vertical scale = cm. Horizontal scale: Total pollen = 100%.

## Conclusions

From the discussion above it is apparent that there have been major changes in the form of the Start Bay coastline. During the early Holocene the shoreline rapidly traversed the low relief to the present offshore zone and under a macrotidal regime this developed a system of salt marshes, estuaries and ephemeral lagoons. Evidence of relict barriers is scanty and it seems unlikely that substantial structures were formed until the shoreline was close to its present position around 5,000 B.P. and the rate of transgression declined. It has not been possible to establish whether the process was continuous or stepwise as envisaged by Rampino and Sanders (1981). The decline in the rate of transgression is thought to result both from a reduction in the rate of eustatic change and to the fact that the coast lies close to a relict cliffline of pre-Holocene age. Major accumulations of gravel are restricted to sections where rockhead is below modem sea level and overwashing can spread gravel across the surface of Holocene valley fills. There is a space problem and only in the Slapton embayment is there room for further movement without a substantial eustatic rise.

It appears that gravel is being withdrawn from other sections of Start Bay to nourish Slapton Sands and enable its crest to keep pace with a rising sea level. This growth will cause a local increase in the amount of sediment that has to be moved in order to bring about shoreline recession and also makes it more difficult for tidal action to maintain residual inlets to a relatively small basin.

As the relief of the sea bed beyond the present coastline is relatively small, gravel may well have been more evenly distributed along the shore in former times. In the circumstances it seems that Slapton Sands act as a sink to which an increasing proportion of the material gravitates as the barrier moves inland and stabilises. Thus the cliff-line not only forms a physical obstacle to further onshore movement but plays a fundamental role in determining the present and future disposition of gravel on the shoreline.

The history of Slapton Sands is particularly interesting. They represent the largest accumulation of gravel in the Start Bay system and yet occupy the site of a former lagoon and estuary that remained open to the sea until comparatively recent times. During the middle Holocene proto-barriers or spits developed in the north and south of Start Bay linked to the submerging interfluves. These barriers were separated by a tidal feature at Slapton which persisted until about 3,000 B.P. At this time it was no longer possible for a tidal link to the coastal valleys to be maintained, the lagoon was sealed and the back barrier environment switched from saline to fresh water. The reasons for closure are not clearly resolved and are likely to be complex. Hails *(op. cit.)* has argued that the hydrodynamics of Start Bay provide minimum energy levels along its central sector and that this, combined with the declining rate of eustatic change, would make gravels less mobile. The author believes that the physical constraints

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constraints of coastal topography combined with the limited sediment budget have played a major part. The lateral movement of barrier material towards the site of former inlets during the late Holocene has also been reported by Halsey (1979) and Oertel (1979) from work on the Atlantic coasts of Delaware and Georgia respectively. In each case relict topography has played a critical role in determining the course of barrier evolution. Thus, although local factors are significant, the course of events in Start Bay appears to conform with a general model which may be widely applicable in the study of such shorelines.

Evidence from the back barrier lagoons has allowed a detailed reconstruction of events during the 3,000 year period since closure. Later events have increased the height of the barrier and produced washover fans, notably during the first four centuries A.D. - a period during which beach movements took place at many sites in Southern Britain. At present lake and marsh surfaces are above mean sea level and, as the base level for local catchments, exert a measure of control upon environments and human activity well inland from the present coastline.

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# A guide to the structure of the Lower to Middle Devonian Staddon Grits and Jennycliff Slates on the east side of Plymouth Sound, Devon

TIMOTHY J. CHAPMAN

Chapman, T.J., A guide to the structure of the Lower to Middle Devonian Staddon Grits and Jennycliff Slates on the east side of Plymouth Sound, Devon. *Proc. Ussher Soc., 5, 460464.* Detailed structural cross-sections are presented from the east side of Plymouth Sound. These display a sequence of sedimentary rocks deformed by overfolding and thrusting towards the north-west. The structures are described and brief comments on their origin presented.
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## Introduction

The aim of this paper is to describe the structure of the Lower to Middle Devonian Staddon Grits and Jennycliff Slates on the east side of Plymouth Sound. This will be done by presenting detailed cliff sections. The sections can be used for excursions and may be used in conjunction with C. Pound's paper (this volume) on the sedimentology of these rocks. Although mainly descriptive, brief postulations about the origins of some of the structures are given. The sections were studied as the Ussher Society field excursion on the 6th January 1983.

Although previous work has been published on the structure of the coastal sections on the east side of Plymouth Sound (Ussher 1907, Fyson 1962, Hobson 1976a,b, 1978), and these accounts include recognition of overturned folds, no attention has been paid to the importance of thrusts. The sections presented here are more detailed than any previous and include new observations. According to Hobson (1976a,b) the section lies just to the north of the axis of the Dartmouth Antiform. Recent work in the Torbay area (Coward and McClay, 1983) and the Plymouth area, Chapman *et* al(in press) has shown this to be a complex thrust and related fold structure. The cross-sections have been traced directly from a photo mosaic of the cliffs taken from a boat (compiled by D.M. Hobson) and there is little control on scale and direction. The detailed geology has been inserted from onshore observations, mainly from the foreshore and cliffs, but also from isolated outcrops on the steep vegetated slopes higher up. The approximate line of the sections is shown on Figure 1.

## Cross-section a-b-c (Fig. 2)

This cross-section covers the southern and lower part of the Jennycliff Slates and the northern and topmost part, of the Staddon Grits. The section is characterised by overturned folds and many of the beds at beach level are inverted, as demonstrated by sedimentary structures (slightly graded units and cross-bedding). Over large
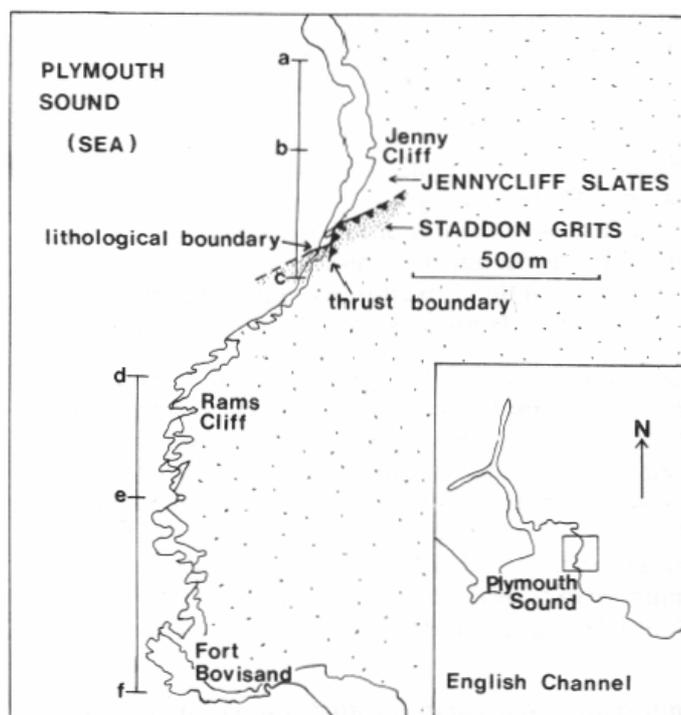


Figure 1. Location map showing the lines of the cliff sections a-b-c (Fig. 2) and d-e-f (Fig. 3).

parts of the section a nearly flat-lying cleavage is less steep than bedding and this is the case for both normal and inverted limbs of the overturned folds, because they are generally nearly recumbent. At locality (1) (grid ref 49145228) a large thrust is evident after inspection of the isolated sandstone outcrop on the beach, and in the undergrowth higher up. Here the flat-lying sandstone is thrust over inverted slates. This thrust, which is off-set slightly by a late normal fault must have a displacement in excess of 30m. One may speculate whether or not this thrust passes down to a basal floor thrust or dies out
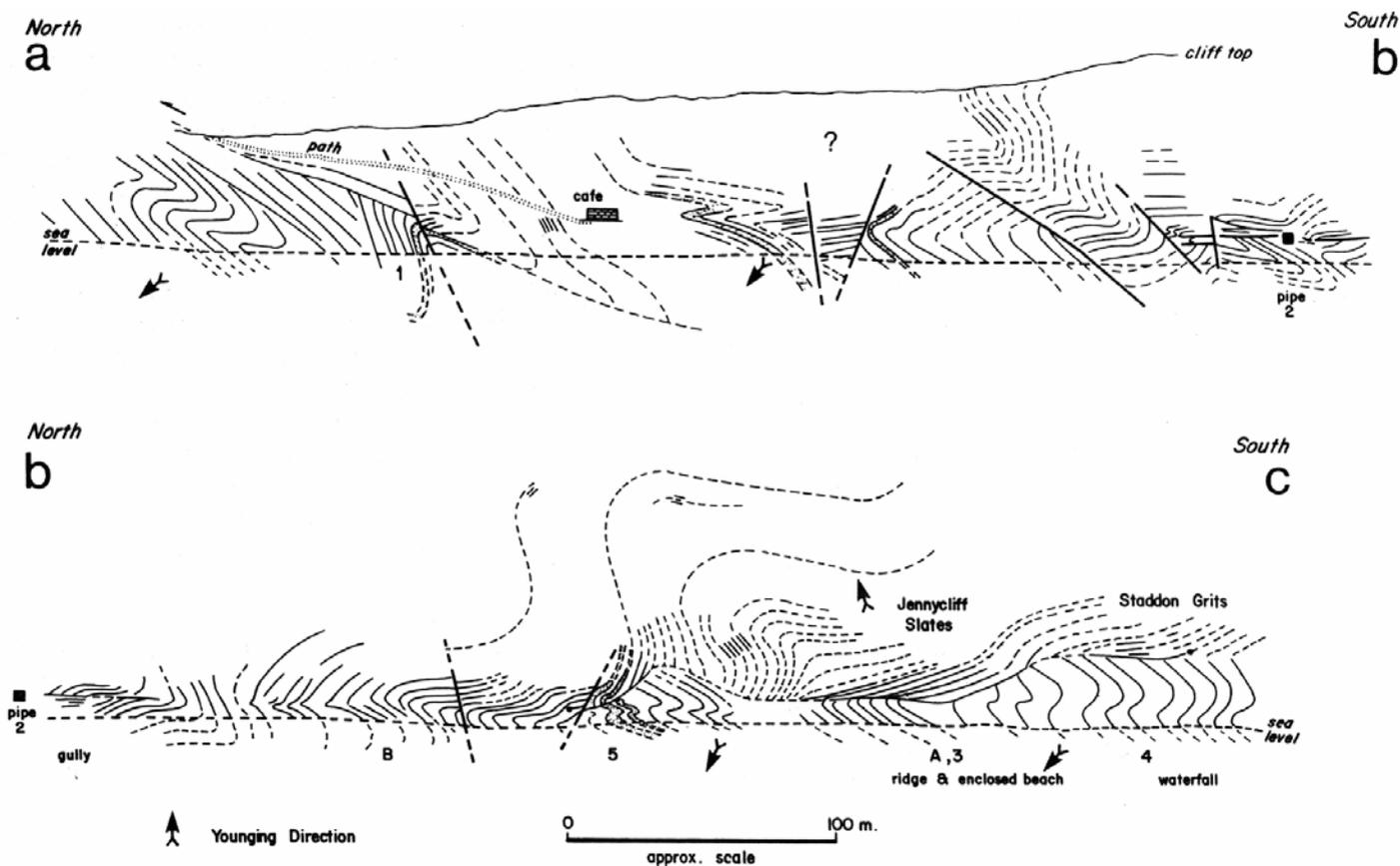
Figure 2. Cliff section a-b-c along Jennycliff and adjacent coast. Rocks are slates with sandstones (stippled). Most of the section is Jennycliff Slates. Capital letters refer to the localities mentioned in C. Pound's paper in this volume. Numbers refer to localities referred to in this paper.

downwards. On this part of the section (Fig. 1) the thrust has placed a backlimb in the hanging wall over a forelimb in the footwall exactly as described by Butler (1982).

At locality (2) on the east side of the pipe (grid ref. 491752093 a medium scale fold-thrust structure is exposed. Fibrous quartz slickensides on the thrust plane indicate overthrusting towards the north west. North of the pipe are several complex small scale fold-thrust structures.

At the southern part of the section between localities (5) and (4) a major flat-lying thrust is exposed. This thrust carries the junction between the higher Jennycliff Slates and lower Staddon Grits northwards although the amount of translation is exaggerated on the cross-section because it is oblique to the slip direction. This structure is best studied starting at locality (3) then moving to (4) and *then* back to (5). This end of the section can only be studied about an hour-and-a-half on each side of low tide.

The thrust is clearly visible at locality (3) (grid ref. 49075186) where Staddon Grits, dipping gently northwards, lie on vertical to inverted Jennycliff Slates (including thin crinoidal limestones). A zone of brecciated rock lies along the thrust plane. Going southwards the thrust rises up the cliff and disappears beneath vegetation but becomes visible again up in the cliff above locality (4) (grid ref. 49055183). Also of interest at locality (4) are excellent examples of non-cylindrical minor folds, *in situ* in the vicinity of the Staddon Grits/Jennycliff Slates boundary, and on a large displaced slab of rock (mentioned by Hobson 1978). The boundary itself occurs where sandstones form 50% or more of the total rock to the south.

The northern end of the major thrust occurs around locality (5) (grid ref. 49125916) although it is not immediately obvious. Excellent northward facing 'Z' folds (southwards verging) are developed in the inverted beds beneath. One thin folded sandstone in the footwall

has a counterpart in the hanging wall of the thrust and indicates only a small displacement here. Further north no thrust occurs.

## Interpretation

It is possible to speculate about the southern part of this thrust above locality (4). By considering the relative geometry of the right-way up beds above, and the mainly wrong-way up beds below, it is possible to show that the thrust must die out just southwards, as shown on Figure 1. A displacement-distance diagram (Williams and Chapman in press, Chapman and Williams in press) was plotted and this locates the point where the slip on the thrust would die out (although the fracture may continue). The end of the section is exposed along strike, just to the south-west at a less accessible locality (grid ref. 48995174). Here the fold profile is the same but no thrust occurs, confirming the above postulation.

If this postulation is correct then the thrust is of finite length terminating both northwards and southwards. At the southern end the beds above cannot he matched with those below on restoration of the thrust. They occupy about twice the distance in the hanging wall of the same beds in the footwall, giving a relative stretch of about 0.5 (defined by Williams and Chapman, in press). Such a situation can only occur if the thrust propagated southwards through a growing fold (see the Models of Eby *et al,* 1923, Williams and Chapman, in press and Chapman and Williams, in press). The sense of overthrusting is still to the north. This thrust may have nucleated at about its mid-point and then grown, propagating both southwards towards the hinterland, and northwards towards the foreland. Slip was towards the north (actually north-west) at all times and the folds grew in response to slip occurring at the same time as thrust propagation (Williams and Chapman, in press).

## Cross section d-e-f (Fig. 3)

The only part of this section that is easily accessible is the southern part just north of the diving centre at Fort Bovisand. Ramscliff at the extreme north is reasonably accessible, but the intervening parts are very difficult to get to even at low tide. The structures are well displayed from a boat.

The southern half of the section is the inverted limb of a major fold which to the south of the section (south of Bovisand Bay) brings in the lower Meadfoot Group in an anticlinal core (Hobson 1976a, Chapman *et al,* in press). The northern half of the section is the flat limb of a major syncline. Superimposed on both limbs are parasitic folds which are still of a relatively large scale (see section Fig. 3) and, disharmonically with these in terms of wavelength are smaller scale parasitic folds, most of which are associated with thrusts (fold-thrust structures). Sketches of some of these structures are shown in Figures 5 and 6. The two fold-thrust structures at Ramscliff (locality 6,

grid ref. 48655131) both show slip towards the north-west as indicated by fibrous quartz-slickensides. These two fold-thrust structures are close together but separated by a late normal fault. In one case (Fig. 5a) a geometrically necessary fold is developed above a small ramp, adjacent to an overturned syncline. This is an excellent example of an out-of-the-syncline thrust as described by Dahlstrom (1970) and Butler (1982). The more southerly fold-thrust structure is a typical overturned fold lying on a thrust (Fig. 5b). The latter cuts up stratigraphic section and then becomes bedding parallel. A similar structure occurs at Hope's Nose, Torquay, and Williams and Chapman (in press) have argued that the fold grew in front of a propagating thrust which then cut through it. The displacement would probably die out northwards in common with other examples of this type of structure but limpet cover on the rocks below in the footwall prevents an accurate correlation across the thrust. A similar structure is seen at (grid ref. 48605109) on the cross-section although this is not easily accessible: The out-of-the-syncline thrust (Fig. 5a) is probably developed within the fold pair, entirely within the hanging wall of the main thrust (Fig. 5b) as shown in Fig. 5c.

On the south of the section at locality 7 (grid ref. 48705088), and shown in detail on Figure 4, the sequence is inverted and a number upside-down fold-thrust structures are developed. Two of these are large and a number of minor ones are also developed.

The thrusts all display the same sense of slip, which give effectively normal displacements. However, these thrusts and associated folds do accommodate contraction along the bedding. In common with many fold-thrust structures the thrusts are relatively planar and cut across the bedding and then become bedding parallel. This is due to change in orientation of the bedding, rather than the thrust. However, the largest fold-thrust structure at this locality can be described as an upside-down fold-thrust structure with a partial ramp. The partial ramp may be due to the presence of a very thick sandstone member (stippled in Fig. 4). The fold here probably grew in front of a planar propagating thrust which cut down to a lower (stratigraphically higher) level. The ramp is not the cause of the structure however because only one bed is effected by it, but many are cut-out. The slip on these thrusts will die out upwards and one minor fold-thrust actually terminates (Fig. 6), The example at the west side of Fig. 4 is much gentler. This may have resulted from a more rapid propagation of thrust relative to slip (Williams and Chapman, in press and Chapman and Williams, in press) compared to the other fold-thrust structures at this locality.

## Interpretation

There are two possible interpretations for these inverted fold-thrust structures. They may have developed in right-way-up-beds prior to inversion and therefore predate the major inversion. In this case they would have been back-

North

d

cliff top

South

e

Ramscliff

6

sea level

North

e

South

f

Fort
Bovisand

sea
level

7

see Fig. 4

Younging Direction

0          100 m.

approx. scale
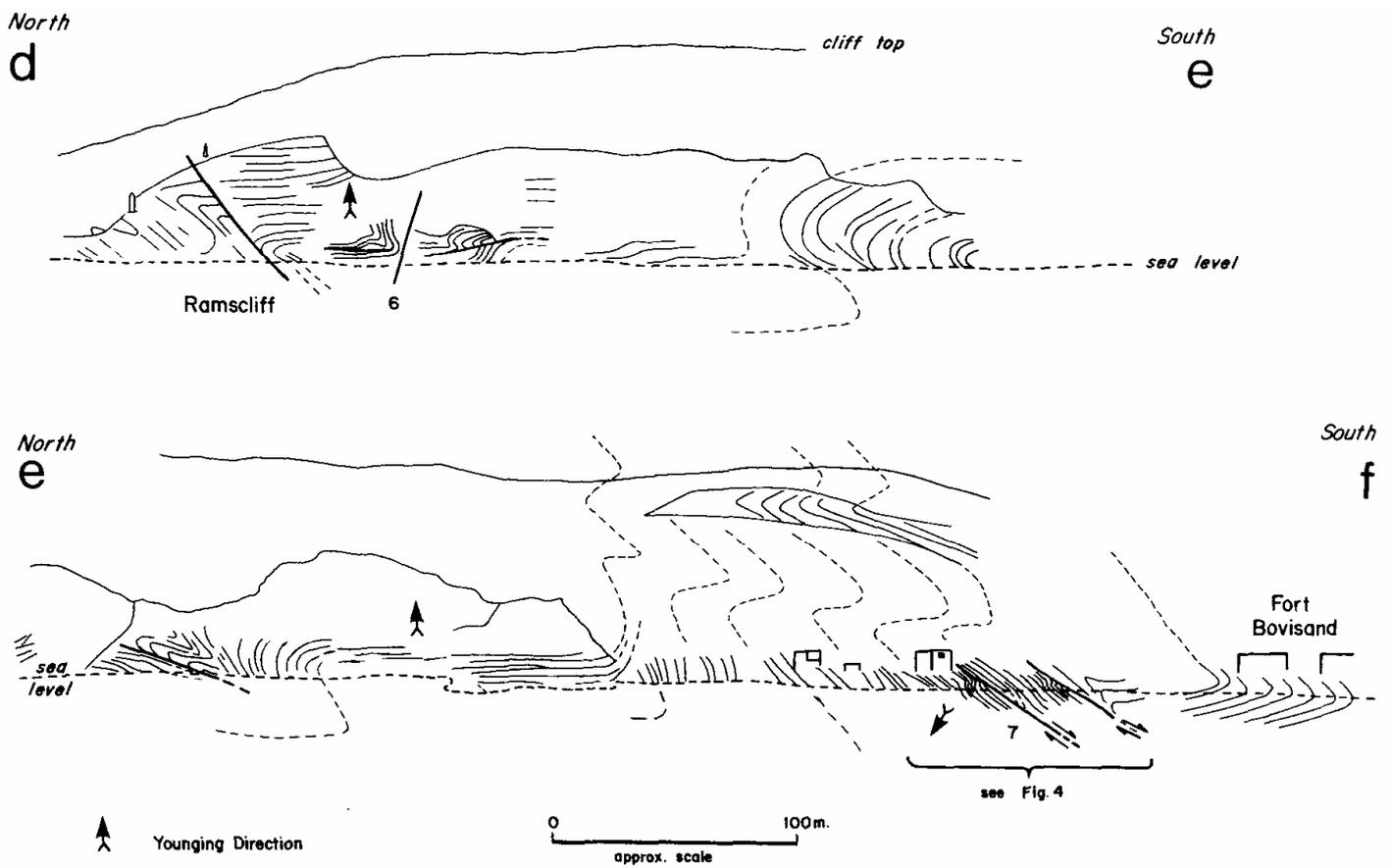
Figure 3. Cliff section d-e-f from Ramscliff to Fort Bovisand.
The section is entirely within the Staddon Grits. Numbers are
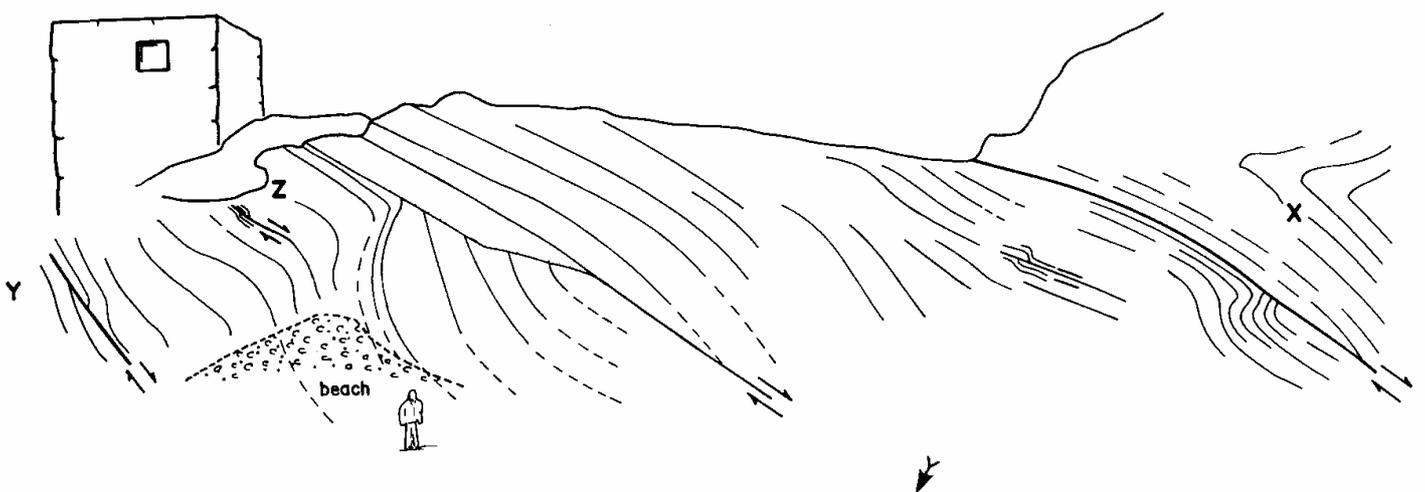localities referred to in this paper.

Z

X

Y

beach

Figure 4. Close up of locality 7, Fig. 3 compiled from
photographs. The stippled bed is a prominent thick sandstone. All
the beds are inverted. The small scale fold-thrust structure
marked z is shown in detail in Fig. 6X and Y are referred to in C.
Pound's paper.

Figure 6. Small-scale fold thrust structure from locality 7 (marked Z on Fig. 4).
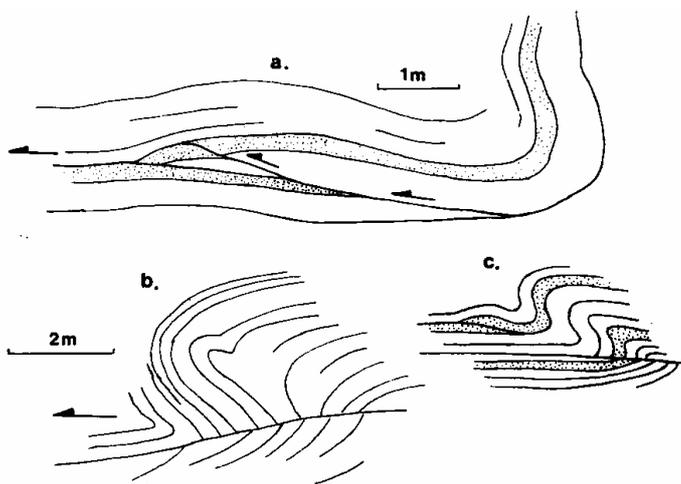
Figure 5. The two fold-thrust structures at Ramscliff (locality 6 section d-e-f) Fig. 5(a) is an out-of-the-syncline thrust which may have developed in the fold pair above the thrust shown in (b). (c) shows probable relationship between (a) and (b).

thrusts with hangingwall slip to the south. All folds, however, are primary with an axial planar first phase penetrative foliation. The more likely interpretation is that they formed in their present orientation in response to flexural slip (Ramsay 1967, 1974) on the inverted limb of a major fold. Flexural slip is indicated by fibrous quartz slickensides on many bedding surfaces. They are therefore parasitic folds and could be true "drag" folds but with associated thrusts. This is certainly implied by the sense of thrusting. If this is the case then the thrusts may die out downwards as well as upwards, as indicated in Figure 7.

Figure 7 shows how these fold-thrust structures and those further north on this section could have formed by flexural-slip on the steep inverted and flat right-way-up limbs of the major folds.
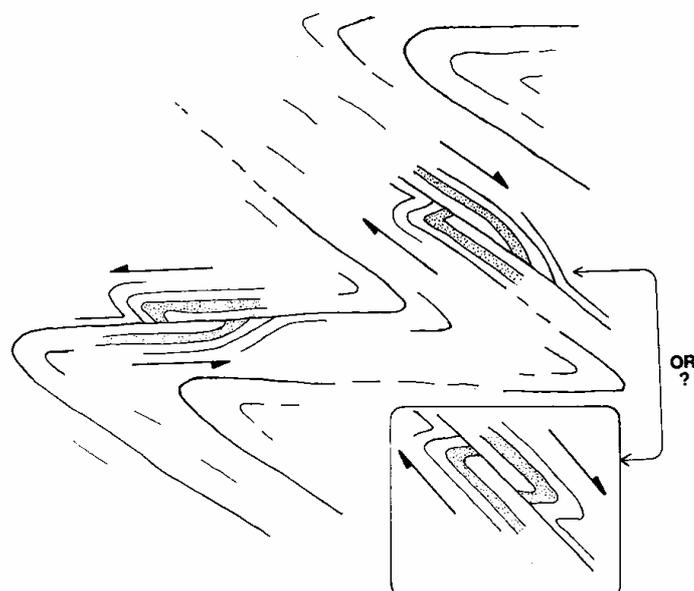


Figure 7. A possible model for the development of normal and inverted fold-thrust structures by flexural slip on the limbs of large scale overturned folds.

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# The sedimentology of the Lower-Middle Devonian Staddon Grits and Jennycliff Slates on the east side of Plymouth Sound, Devon

C.J. POUND

Pound, C.J, 1983. The sedimentology of the Lower-Middle Devonian Staddon Grits and Jennycliff Slates on the east side of Plymouth Sound, Devon. *Proc. Ussher Soc., 5,* 465-472.

In the Plymouth area the Staddon Grits (Emsian) represent the development of a fluvial-dominated, low wave-energy delta. The supply of sediment to the delta ceased in late Emsian times and uppermost Staddon Grits sediments were locally reworked by waves into a series of bars. The succeeding Jennycliff Slates (late Emsian-early Eifelian) record deposition on a storm-dominated shelf. The Staddon Grits are thought to represent a regressive sediment pulse supplied by a local fault block to the north. This source was active throughout the Upper Palaeozoic and may have formed part of a series of east-west trending synsedimentary fault zones, possible in a strike-slip orogenic setting.
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## Introduction

The Lower to Middle Devonian succession along the eastern side of Plymouth Sound (fig. 1) comprises the predominantly continental Dartmouth Beds (Dineley 1966) succeeded by the shallow marine Meadfoot Group (Harwood 1976). The Meadfoot Group consists here of the Bovisand Beds followed by the type section of the Staddon Grits (Holl 1868). Evans 0980) collected brachiopods from lower horizons in the Meadfoot Group along the coast south-east of Plymouth which yielded a fauna diagnostic of a mid-Siegenian to early Emsian age. No fauna has been recorded in the highest Staddon Grits which are probably late Emsian in age (Evans, *pets. comm.).* The Meadfoot Group is succeeded by the shallow marine Jennycliff Slates (Ussher 1907) of probable late Emsian to Eifelian age (Orchard 1977).

This paper describes three logged sections which are representative of facies developments within the Staddon Grits and Jennycliff Slates. Each logged section will be described and interpreted in turn, with their implications for regional palaeogeography being discussed at the end of the paper.

## North Fort Bovisand (SX 48715084)

An inverted sequence from the central parts of the Staddon Grits is well exposed 120m north of Fort Bovisand (fig. 1, location A), where the sequence is cut by a series of bedding-parallel thrusts across which bedding may be matched. The line of logged section (marked X-Y on Chapman's fig. 4, 1983) may be divided into an association of six facies within an overall coarsening- and thickening - upwards sequence which represents the progradation of a delta mouth bar. Each facies is described below and may be located on the log (fig. 2) in the left-hand column.
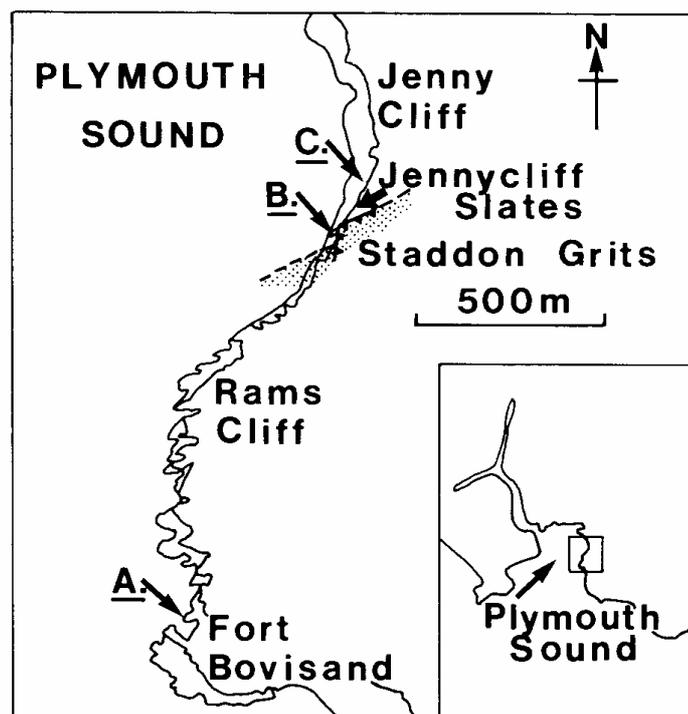


Figure 1. Map showing location of sections described in the text.

## A. Lenticular bedded sandstone and mudstone

This facies consists of silty-mudstone with thin (0.5-1.5cm) unconnected and connected fine sandstone lenses, the sandstones consisting of form-discordant, bipolar, cross-laminations in beds 5-60cm thick, Suggestive of storm-wave generation with mud deposition during fair weather.

## B. Sand-streaked mudstones

Horizontally-laminated sand-streaked mudstones, 0.2-2cm in thickness characterise this facies. Two small slumped horizons occur in this facies which display poorly defined folds and listric micro-faults and indicate a palaeoslope dipping to the SSW. Thick sand-streaked mud facies of this type commonly occur in delta front sequences, the coarser beds representing flood-generated sediment incursions from the distributary mouth (eg Coleman, 1976).

## C. Flaser bedded sandstone

These flaser beds consist of l-3cm thick sets of bipolar cross-laminations which display features diagnostic of wave generation, forming cosets up to 40cm in thickness. Mud flasers are frequently preserved in ripple troughs and, occasionally, mud laminae drape ripple crests. This facies represents wave-generated shoals developing in an environment of fluctuating wave energy, the muds draping the ripples during periods of low wave energy. *D. Current cross-laminated sandstones*

In this facies, cosets up to 25cm in thickness, are composed of 1-2cm thick sets of current- and climbing-rippled fine sandstones. The current rippled cosets indicate transport directions to the south and are interpreted as the products of river-flood processes supplying high sand concentrations to the distal mouth bar *(cf* Coleman, 1976).

## E. Parallel-laminated sandstones

In this facies, 5-70cm thick beds of parallel-laminated fine sandstones are separated by thin mudstone laminae. Typically, the bases of units are scoured, frequently with basal intraformational clasts of mudstone or siltstone; the latter are also found scattered throughout units. These thick parallel-laminated sandstone units are interpreted as the rapidly deposited sediments of river-flood Currents with substantial sediment loads *(c.f.* Coleman, 1976).

## F. Massive-appearing and cross-bedded sandstones

These thick, often massive-appearing, sandstones infill a major channel form (shown in fig. 3) with faint parallel laminations picked out locally. This channel form incises into a 20cm thick set of planar cross-bedded sandstone. Further planar and trough cross-bedded sandstones occur above the massive-appearing channel fill. The cross-beds indicate a transport direction to the south. Thin sections from the channel infill reveal moderately sorted lithic arenites composed of strained metamorphic quartz (60%), crystal tuffs (20%), dust tuffs (6%), quartzite (8%), indeterminate plagioclase (4%), detrital muscovite (1%) and heavy minerals, predominantly rutile, (1%).

This facies represents a minor distributary channel sequence with the cross-bedded sand-stones reflecting the downstream migration of dunes.

*Sequences*

Field examination by the author revealed that, except for the highest 25m (discussed in the following section), the 400m thick succession of Staddon Grits along the eastern side of Plymouth Sound is composed of a series of coarsening-(CU) and thickening-upwards (TkU) sequences. The described sequence north of Fort Bovisand occurs 150m above the base of the Staddon Grits and provides the best exposed example of a CU, TkU sequence. These upwards coarsening and thickening sequences reflect a gradual upwards increase in energy through the sequence with a passage from offshore marine to nearshore conditions indicated. Such a sequence can be matched with many deltaic settings *(eg* Kelling and George 197 I; Elliott 1976), passing from low energy delta front to mouth bar and channel environments (cf Coleman 1976; Elliott 1978).

The base of the logged sequence occurs 2m above a major sandstone body that marks the top of the preceding CU, TkU sequence. In the lower part of the logged succession the transition from current cross-laminated (facies D) and flaser bedded sandstones (facies C), interbedded with lenticular beds (facies A) to thick parallel-laminated sandstones (facies E) represents a distal to proximal mouth bar sequence. At 7.5m on the log a minor CU sequence is initiated where the succession becomes more muddy, possibly reflecting the abandonment of a nearby distributary channel. Waves were minimal, only locally reworking sands deposited by fluvial processes. This resulted in a wave-generated bar being produced on the proximal mouth bar (11.5m on the log). Terminating the cycle is a distributary channel sequence (facies F) followed by a series of parallel-laminated siltstones and sandstones. The latter continue for 4m above the top of the logged sequence and are abruptly replaced by thick mudstones at the base of the next major cycle. The parallel-laminated units are poorly exposed, but may represent a crevasse splay or transgressive sequence.

Applying the scheme of Galloway (1975) to the Staddon Grits the deltaic complex would be defined as a fluvial dominated, low wave-energy system. The fetch of the basin into which this delta prograded may be *estimated* from the measurements of wave-ripple profiles from the succession described above. Employing the methods of Allen (1979) to the limited measurements (N=7), a wave period of 5.5 seconds was calculated for the storm-produced ripples at the threshold condition (vertical form index = 8.33). For storm conditions (winds 30-40 knots), where fetch distance is limiting, the graph of Neumann (1953) indicates a fetch *in the order of* 70km.

Figure 2. North Fort Bovisand (SX 48715084): Log through the central Staddon Grits. Facies: A = Lenticular bedded sandstone and mudstone; B = Sand-streaked mudstones; C = Flaser bedded sandstone; D = Current cross-laminated sandstones; E = Parallel-laminated sandstones; F = Massive-appearing and cross-bedded sandstones.

METRES

FACIES
A B C D E F

MAX.
GRAIN
SIZE

MUD
SILT
V. F. Sd.
F. Sd.
M. Sd.

PERCENT
SAND
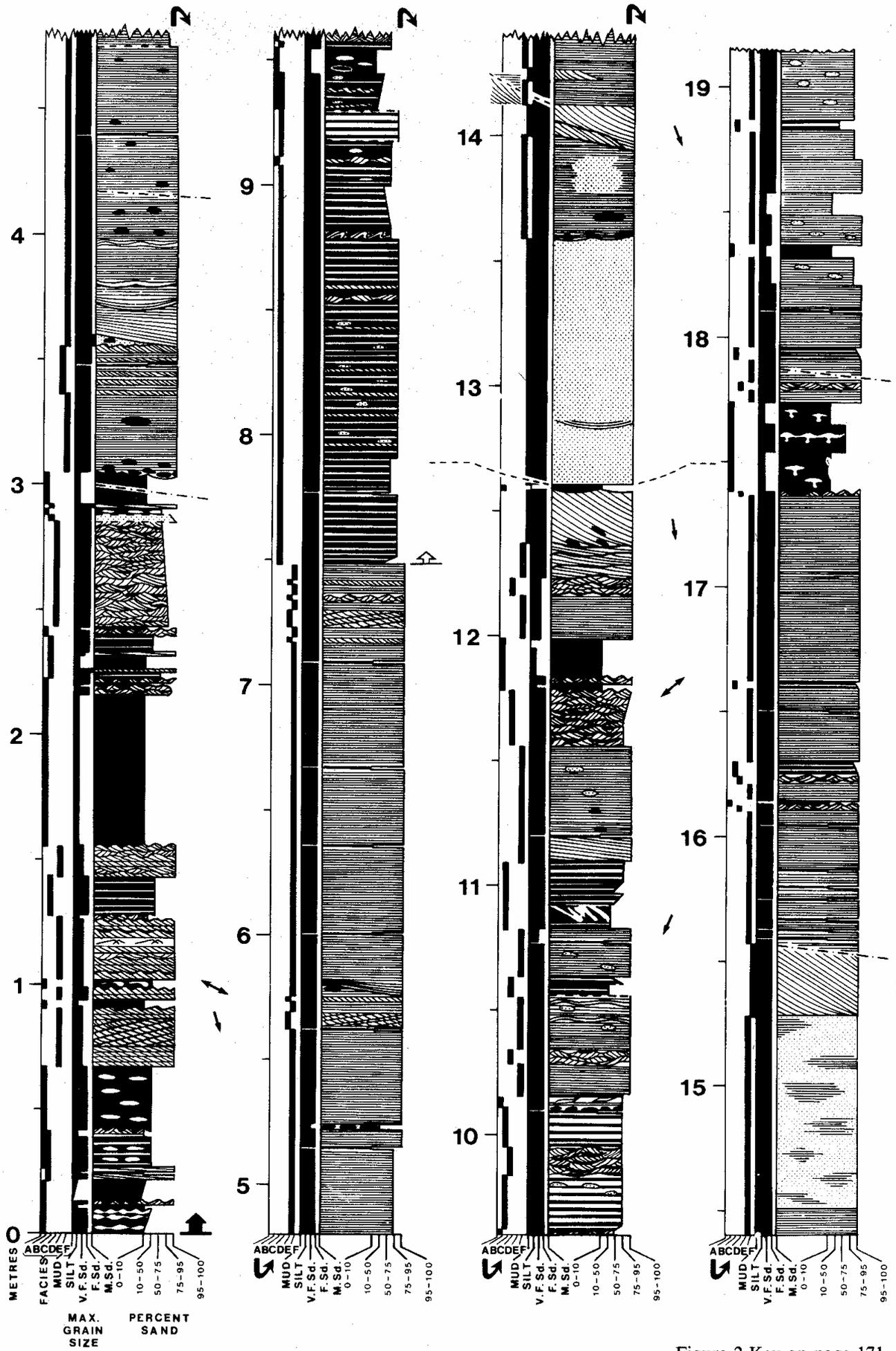
0 - 10
10 - 50
50 - 75
75 - 95
95 - 100

Figure 2 Key on page 171.

Figure 3. North Fort Bovisand (SX 48715084); Distributary channel facies (facies F). Major channel-form incising into southerly migrating (to the right in the photograph) straight-crested dunes. Photograph from an inverted sequence. Bar scale 50cm. in length.

## Jennycliff Bay: Staddon Grits thrust above Jennycliff Slates (SX 490,75185)

This locality (marked B on fig. 1; A on fig. 2 of Chapman, 1983) displays gently dipping uppermost Staddon Grits thrust over younger, inverted, Jennycliff Slates. The upper 25m of Staddon Grits consist of a series of 2m thick sandstone bodies alternating with heterolithic sediments which pass conformably into the overlying Jennycliff Slates. The log (fig. 4) shows the Staddon Grits immediately below the Jennycliff Slates. The sequence is divisible into three facies which are shown in the left-hand column of the log.

### A. Lenticular bedded sandstones and mudstones

The lenticular bedded members (Reineck and Wundefiich, 1968) consist of thin sandstones (l-4cm thick) which are cross-laminated and have a parallel/low-angle laminated base. The tops of these sandstones display rounded, or occasionally trochoidal, rippled profiles with a cross-lamination style characteristic of generation by waves (Boersma, *In:* de Raaf *et al.* 1977). Wave ripple crests trend E-W. The heterolithic nature of this facies reflects alternations in energy, the thin sandstones representing a temporary lowering of wave base during storms.

### B. Flaser bedded sandstone and mudstone

Flaser bedding (Reineck and Wunderlich, 1968) occurs with a cross-lamination (see base fig. 5) which is diagnostic of wave-generated structures. Locally, the sandstone takes on a more massive appearance with only isolated patches of wave generated lamination. These massive-appearing sandstones are quartz cemented, well sorted sublitharenites and are therefore more mature than the lithic arenites exposed in the Bovisand section.

The flaser beds represent an environment of more constant energy conditions than the lenticular beds, the massive sandstones indicating a constantly agitated environment, the substrate having built above fair weather wave base.

### C. Parallel-laminated and hummocky cross-stratified sandstones

In this facies, amalgamated, massive and parallel-laminated sandstone units are separated by thin, silty-mudstone laminae. In addition, some units display low-angle, concave- and convex-up laminae and set intersections characteristic of hummocky cross-stratification(Harms 1975).

The suite of structures displayed by this facies is indicative of storm-generated sandstones deposited between storm and fair weather wave base.

### Sequences

The above three facies define a series of sandstone bodies which represent sandy shoals deposited on a muddy shelf, similar to those described by de *Raafet al* (1977) from the Lower Carboniferous of County Cork, Ireland.

The basal 0.6m of the log shows the top of a sandstone body consisting of amalgamated storm sandstones which represent a storm-generated shoal deposited between storm and fair weather wave base.

The bulk of the logged succession (0.6-3.35m on the log) is composed of a symmetrical development of facies coarsening towards the centre of the unit and then fining upwards (CUFU sequence of de Raaf *et al,* 1977). This sequence is interpreted as a section through the margin of a submerged bar which tapers out and digitates laterally (cf de Raaf *et al,* 1979 fig. 23).

Figure 4. Jennycliff Bay, Staddon Grits above thrust (SX 49075185), and Jennycliff Slates (SX 49075186). Logs through uppermost Staddon Grits and basal Jennycliff Slates Facies A = Lenticular bedded sandstones and mudstones; B = Flaser bedded sandstone; C = Parallel-laminated and hummocky cross-stratified sandstones.

Figure 4 Key on page 171.

METRES · FACIES · A B C · BIOTURBATION · PERCENT SAND · 0–10 · 10–50 · 50–75 · 75–95 · 95–100

STADDON GRITS

A B C · 0–10 · 10–50 · 50–75 · 75–95 · 95–100

METRES · 0–10 · 10–50 · 50–75 · 75–95 · 95–100
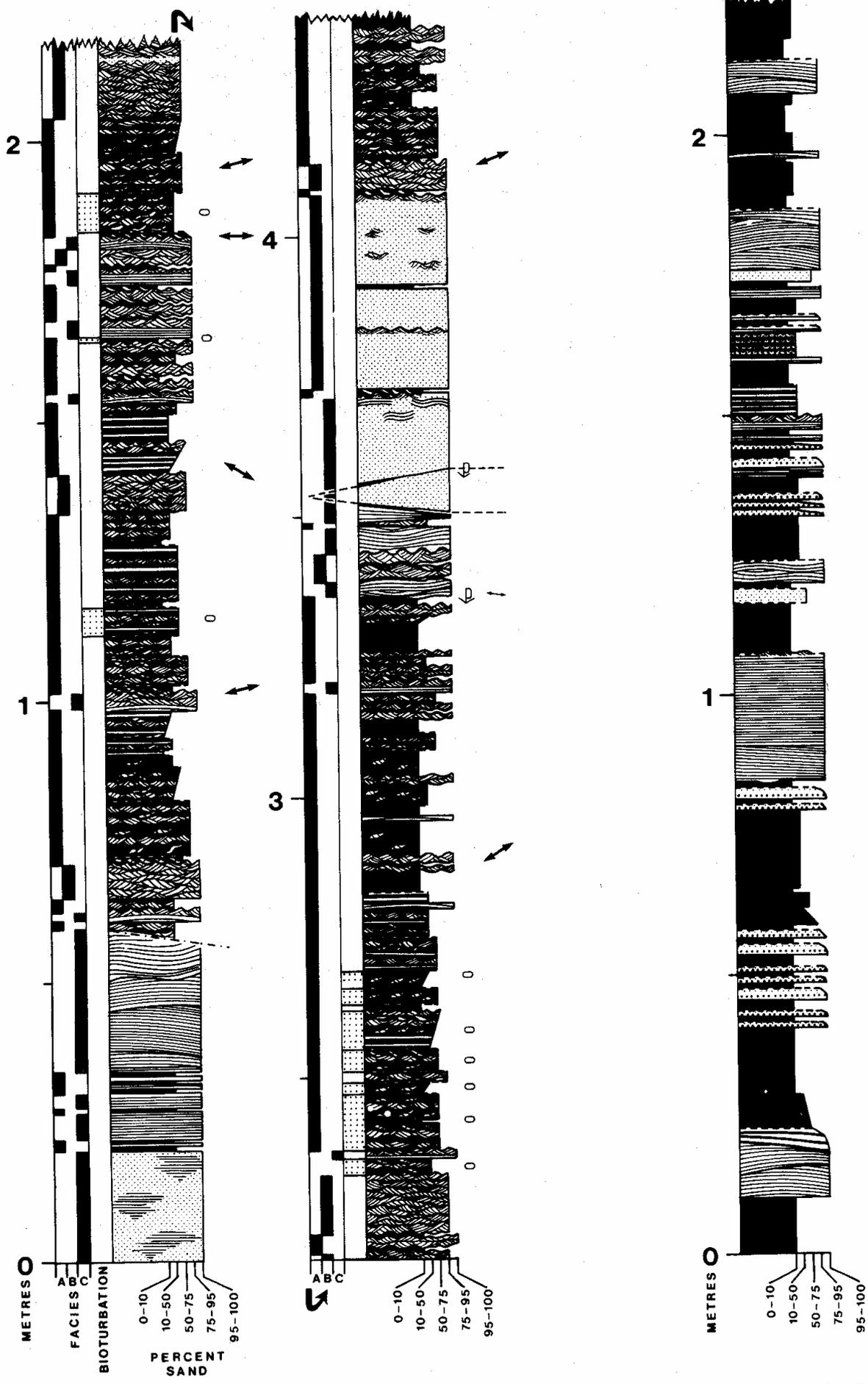
JENNYCLIFF SLATES

Figure 5. Jennycliff Bay Staddon Grits above thrust (SX 49075185): Basal two-thirds of photograph shows bipolar, wave-produced laminations. Upper one-third shows sandstone streaked silty-mudstones. Scale 3.5cm in length.

Above, a CU sequence with storm-generated sandstones passes upwards into sandstones with a massive appearance. The CU sequence is immediately succeeded by a thin FU sequence which shows wave-flaser passing upwards into lenticular bedding.

The CU sequence with a thin FU top represents a section through the main part of a wave-generated bar *(cf* de Raaf *et al,* 1979, fig. 23). The massive sandstone infers a constantly agitated environment, the bar having built-up above fair weather wave base.

## Jennycliff Bay: Lower Jennycliff Slates (SX 49075186)

This locality (marked C on fig. 1; B on Chapman's fig. 2, 1983) displays an inverted sequence of lithologies typical of the lower Jennycliff Slates.

The logged sequence (fig. 4) occurs 80m above the base of the group and shows a series of silty-mudstones, the silt content of some of the units decreasing upwards, interbedded with thin and thick bedded sandstones. The thinner sandstones (0.5-2cm thick) are normally graded and have sh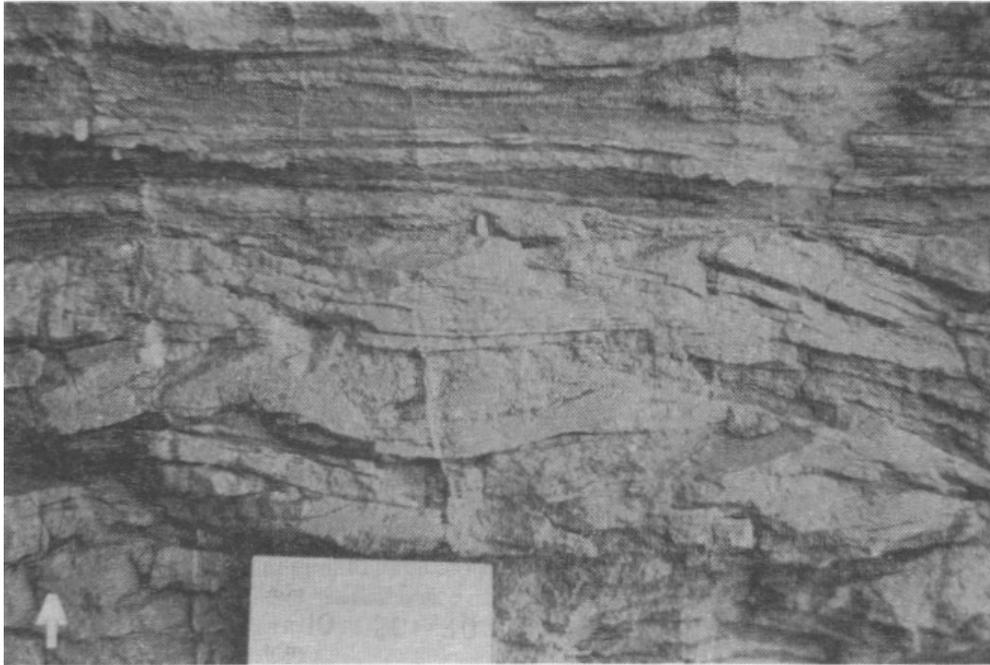arp bases and diffuse tops. Beds of this type were termed graded rhythmites by Reineck and Singh (1972) who noted that they formed from suspension clouds settling-out near storm wave base under the influence of oscillatory currents of minimal energy.

The thicker bedded sandstones (4-23cm thick) are planar based and have gradational, and in one case current rippled, tops. Each unit is 2-3 sets thick, the sandstone

displaying low-angle, convex- and concave-up laminae, set intersections also being curved and at a low angle. With the exception of the unit at lm on the log, which is composed of parallel-laminated sandstone, the thick sandstones exhibit hummocky cross-stratification, a storm-generated structure (Harms 1975).

The basal Jennycliff Slates represent deposition near storm wave base. The graded rhythmites and thin laminated sandstones record deposition during storms, whilst the hummocky cross-stratified sandstones indicate more exceptional storms.

## Discussion and Conclusions

The widespread late Siegenian transgression (House, 1975) resulted in the replacement of the continental Dartmouth Beds (Dineley, 1966) by the shallow marine Meadfoot Group. The latter Group is composed of two formations: the Meadfoot Beds and Staddon Grits. Richter (1967) concluded that the Meadfoot Beds were deposited in fairly shallow water, but not in the intertidal zone as there was no evidence of exposure. A subtidal setting was confirmed by Evans (1980) using brachiopod associations. In contrast, Selwood and Durrance (1982) have proposed a tidal flat and lagoonal environment. A shallow marine setting, as proposed by the former two authors, was confirmed by field examination of exposures in South Devon and Cornwall by the author. However, evidence for tidal activity is equivocal, and a storm-dominated shelf setting seems more consistent with observations.

## LEGEND

| | | | |
|---|---|---|---|
| ▰ | HOMOGENEOUS MUD/SILT | ▨ | PLANAR CROSS-BEDDING |
| ⌇ | MUD LAMINA | ▤ | PARALLEL-LAMINATION |
| ⬮ | SAND LENS | ▨ | HUMMOCKY CROSS-STRATIFICATION |
| ⌁ | ISOLATED, WAVE-PRODUCED LENS | ▨ | SAND:MASSIVE APPEARANCE |
| ▨ | CONNECTED, WAVE-PRODUCED LENS | ᴫᴫ | SLUMPED HORIZON |
| ▼ | LOADED SAND LENS | ⬮ MUD ⎫ | |
| ▬ | SAND STREAK | ⬯ SILT ⎭ INTRAFORMATIONAL CLASTS | |
| ▤ | PARALLEL-LAMINATED SAND STREAK | BASES: | |
| ▬ | LOW-ANGLE LAMINATED SAND STREAK | ▭ | PLANAR |
| ▨ | GRADED RHYTHMITE | ⌁ | GRADATIONAL |
| ▨ | WAVE-RIPPLED ZONE | ⌒ | UNDULATORY |
| ▨ | CURRENT-RIPPLED ZONE | ⌇ | SCOURED |
| ▨ | CLIMBING CURRENT-RIPPLES | ▨ | SCOURED:WITH CONCORDANT INFILL |
| ⌁ | COMBINED-FLOW RIPPLES | ⟷ | DRAG MARKS:WITH ALIGNMENT |
| ⌇ | TROCHOIDAL ⎫ | ⌁ | PROD MARKS |
| ⌇ | STEEP, OPEN ⎬ WAVE-RIPPLED TOPS | ⟶ | CURRENT DIRECTION |
| ⌇ | LOW-AMPLITUDE ⎭ | ⟋ | WAVE-RIPPLE CREST ALIGNMENT |
| ▨ | WAVE-FLASER BEDDING | ▢ | *PLANOLITES*-TYPE BURROWS |
| ▨ | TROUGH-CROSS BEDDING | ⌁⌁⌁ | BEDDING-PARALLEL THRUSTS |

The Staddon Grits were recognised by Simpson (1951), Dineley (1961),Harwood (1976) and Evans (1980) to be a lenticular facies development at the top of the Meadfoot Group. Information on the depositional setting of the Staddon Grits is limited. Harwood (1976) proposed a nearshore setting, a theme expanded upon by Evans (1980) who suggested that conditions were unsuitable for brachiopod colonisation "owing to the influx of large amounts of clastic sediment". However, Selwood and Durrance (1982) envisaged an offshore bar setting, sediment being supplied to the coast by rivers discharging from a Dartmouth Beds alluvial plain type facies.

Evidence presented in this paper suggests that the bulk of the Staddon Grits represent a deltaic incursion of sediment derived from the north. The petrographic assemblage described from the distributary channel sequence (facies F) is significantly different to that of the Dartmouth Beds where "quartz, muscovite, iron ores and some biotite are the most conspicuous grains, making up 85% of the total" (Dineley 1966), arguing against the derivation of the Staddon Grits from Dartmouth Beds type lithologies. During uppermost Emsian times the supply of sediment to the delta ceased and the deltaic deposits were locally reworked into a series of wave-generated bars. Deepening resulted in deposition of the Jennycliff Slates on a storm-dominated shelf below storm wave base.

Selwood and Durrance (1982) proposed a simple transgressive sequence from alluvial plain deposits (Dartmouth Beds) to tidal flats and lagoons (Meadfoot Beds), offshore bars (Staddon Grits) and finally, open marine shelf (Middle Devonian Slates). However, the interpretation of the Staddon Grits as a deltaic incursion of sediment suggests that the above model is over-simplified as the Staddon Grits represent a regressive phase.

Petrological and palaeocurrent evidence suggest a nearby and northerly source for the Staddon Grits, and the idea of Hendricks (1959) is revived, with the Staddon Grits developing on the margin of a Staddon ridge. It is proposed that this feature could explain the petrologically distinct regressive phase of Staddon Grit sedimentation. This feature was drowned by the widespread basal Middle Devonian transgression (House 1975), and sedimentation continued with deposition of the Jennycliff Slates below storm wave base.

Periodic reactivation of the Staddon ridge during the Upper Palaeozoic could explain a number of sedimentation events. Both the Torquay Limestone (Scrutton 1977) and the Plymouth Limestone (Braithwaite 1966) were proposed to have been deposited on a schwelle trending east-west. Ussher (1907) reported petrographic assemblages from the Upper Devonian Wearde Grit north of Plymouth which are similar to the Staddon Grits assemblage. Later, in Lower Carboniferous times, Whiteley (1982) has proposed a high which locally sourced turbidites from the south of the St Mellion area. Each of these facies developments may relate to the reactivation of a fault controlled block.

Matthews (1977) proposed that Upper Palaeozoic sedimentation in the south-West of England was largely controlled by a series of east-west trending fault zones. Synsedimentary movement along fractures of this nature could result in the localized supply of sediment frown ephemeral fault-blocks (*cf* Tunbridge, 1983). Furthermore, the hypothesis that the Hercynides formed in a dominantly strike-slip orogen (*eg* Badham, 1982) seems particularly appropriate, as short-lived fault-uplifts are a characteristic feature of strike-slip motions (Reading, 1980).

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# Late Triassic and earliest Jurassic palynomorph assemblages from the Western English Channel and neighbouring areas

G. WARRINGTON

Warrington, G. 1983. Late Triassic and earliest Jurassic palynomorph assemblages from the Western English Channel and neighbouring areas. *Proc. Ussher Soc., 5,* 473-476.
Late Triassic (Carnian) miospores were recovered from mudstones proved beneath Cretaceous rocks in *Zephyr* well 87/16-1 in the South-Western Approaches Basin. In the western English Channel Basin, Late Triassic (Rhaetian) to Early Jurassic (Hettangian) palynomorph assemblages were obtained from the Lower Lias and underlying beds, identified as the Penarth Group, in *Zephyr* well 88/2-1, and from beds proved beneath *liasicus* or *planorbis* Zone (*johnstoni* Subzone) Lias deposits in the "Sealab" SLS 17 borehole.
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## Introduction

A large number of samples have been examined for microfossils to establish the ages of formations encountered during offshore geological survey work by the Marine Geology Unit of the Institute of Geological Sciences in the Western English Channel area. The results of these studies, many of which were carried out on material obtained by shallow-sampling devices, have been largely summarised by Warrington and Owens (1977) and Wilkinson and Halliwell (1980). Deeper penetration of the offshore successions was achieved in the boreholes of the "Sealab" and *Zephyr* drilling programmes and the results of palynological work carried out on Late Triassic and earliest Jurassic sequences proved in a number of these boreholes in the Western English Channel Basin and adjacent parts of the South-Western Approaches Basin are documented in this account. Depths in the *Zephyr* boreholes are relative to drilling datum (the kelly bushing, KB) and in the "Sealab" borehole are relative to seabed.
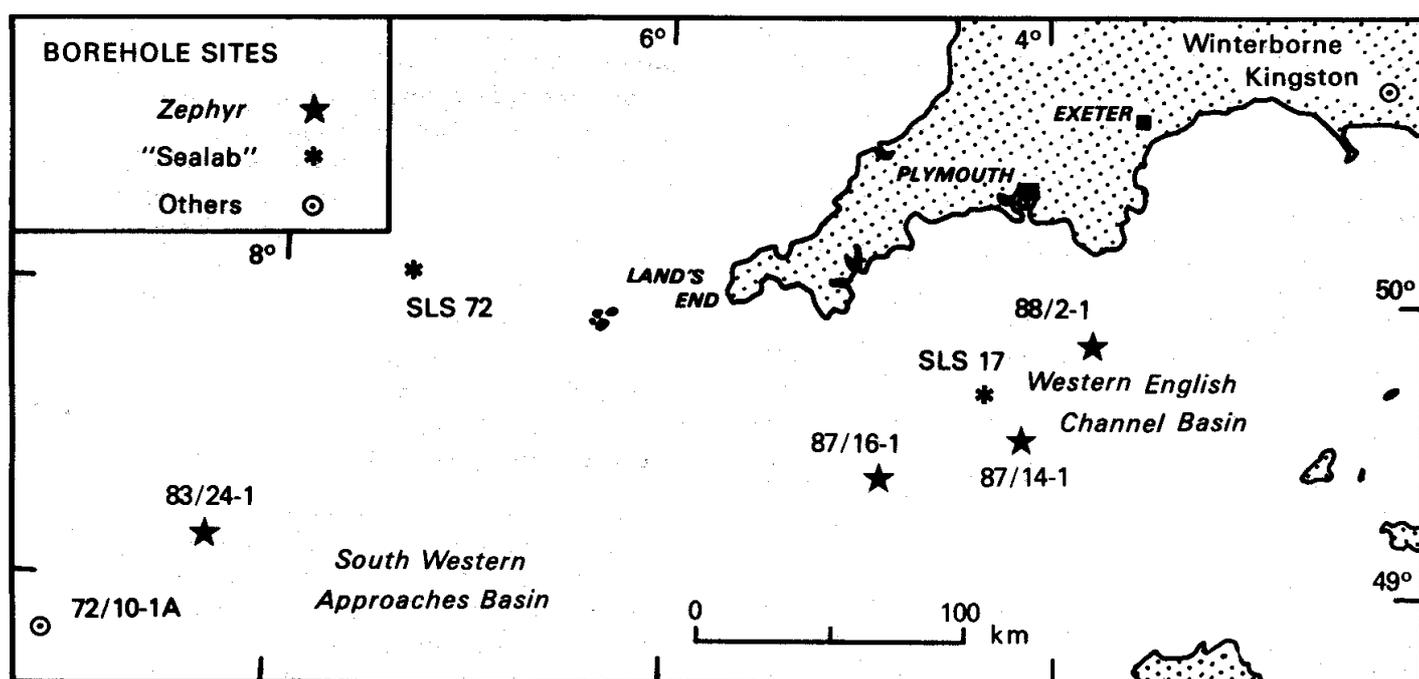


Figure 1. The locations of boreholes referred to in this account.

## Palynological Results

### 1. The Zephyr Boreholes

The *Zephyr* boreholes, designated offshore wells 83/24-1, 87/14-1, 87/16-1 and 88/2-1 (Fig. 1), were drilled in the Western English Channel and South-Western Approaches basins in 1977 (Evans *et al.,* 1981). Palynological samples from red-bed successions of presumed Permo-Triassic age proved in wells 83/24-1 and 87/14-1 were unproductive (Evans *et al.,* 1981, pp.9,18) and those sections are not considered further. The following results were obtained from the remaining *Zephyr* wells.

*Well 87/16-1:* In this well, Albian rocks rest unconformably, at 829.5m, upon mudstones assigned to the Mercia Mudstone Group; these were proved for 100m to the logged terminal depth of 929.5m below KB (Evans *et al.,* 1981, pp.21,22).

Nine samples of well-cuttings representing the mudstone succession below 849m yielded Cretaceous palynomorphs which ore considered to be contaminants derived, by caving, from the overlying Albian and younger Cretaceous rocks. However, in preparation MPZ 1623, from the lowest sample (927-929.5m), these remains occur in association with and are numerically subordinate to moderately well preserved specimens of the following Triassic miospores:

> *Camerosporites secatus* Leschik emend. Scheuring 1978
> *Duplicisporites verrucosus* Leschik emend. Scheuring 1978
> *Ellipsovelatisporites plicatus* Klaus 1960
> *Enzonalasporites vigens* Leschik *1955, sensu* Scheuring 1970
> *Ovalipollis pseudoalatus* (Thiergart) Schuurman 1976
> *Patinasporites densus* Leschik emend. Scheuring 1970
> *Triadispora obscura* Scheuring 1970
> *Vallasporites ignacii* Leschik *1955, sensu* Scheuring 1970

This miospore association is dominated by specimens of *Camerosporites secatus,* and the occurrence of that form with *Patinasporites densus* and *Vallasporites ignacii* is indicative of a Late Triassic, mid- to late Carnian age (cf. Mostler and Scheuring, 1974; Visscher and Krystyn, 1978). The Late Triassic age obtained from this sample is considered applicable to the remainder of the red-brown mudstone sequence proved beneath Albian rocks in this well and thereby supports its assignment to the Mercia Mudstone Group. It is possible that the higher part of that sequence may include beds of post-Carnian (i.e. Norian) age.

*Well 88/2-1:* Lias deposits underlie Cenomanian Chalk in this well and rest conformably, at 705.5m, upon the Penarth Group which succeeds the Mercia Mudstone Group at 753m. The Mercia Mudstone Group, including a representative of the Blue Anchor Formation, was proven for some 120m to the logged terminal depth of 872.9m below KB (Evans *et al.,* 1981, pp.25-27).

No palynomorphs were recovered from the succession proved beneath the Penarth Group; material from 870.7m, in the terminal core, was unproductive. Palynomorph assemblages comprising miospores and organic-walled microplankton, associated with test-linings of foraminifera, were, however, obtained from six samples of well-cuttings representing the basal Lias and Penarth Group succession between 678 and 744m (Fig. 2). These assemblages are divisible into two groups on the basis of the downhole appearances of the dinoflagellate cyst *Rhaetogonyaulax rhaetica* (Sarjeant) Loeblich and Loeblich emend. Harland, Morbey and Sarjeant 1975, and of the miospores *Rhaetipollis germanicus* Schulz 1967, *Ricciisporites tuberculatus* Lundblad 1954, and the first definite specimens of *Ovalipollis pseudoalatus* (Thiergart) Schuurman 1976 at 729m. Assemblages from the two preparations (MPZ 686 and 690) from below that level (Fig. 2) are of Rhaetian age and are comparable with those obtained from the Westbury Formation and Cotham Member (Lilstock Formation) of the Penarth Group proved in the Winterborne Kingston borehole, Dorset (Fig. 1; Warrington 1982). The four assemblages from above 729m (preparations MPZ 669, 675, 679 and 683, from 678-723m: Fig. 2) contrast markedly in composition with those from below that level and are comparable with those obtained from the Langport Member (Lilstock Formation) of the Penarth Group and from the basal Lias in the Winterborne Kingston borehole. An horizon between 723 and 729m in the *Zephyr* well is, therefore, correlatable palynologically with the Cotham Member - Langport Member boundary in the Penarth Group succession of Dorset.

The Triassic-Jurassic system boundary is defined by the lowest occurrence of ammonites of the genus *Psiloceras* and is usually located a few metres above the base of the Lias facies (Cope *et al.,* 1980; Warrington *et al.,* 1980). No change occurs in palynomorph assemblages at a corresponding level and this boundary cannot, therefore, be located in *Zephyr* well 88/2-1 on the basis of the palynological results presented here. However, from a comparison with the Winterborne Kingston borehole (Evans *et al.,* 1981, Fig. 20), the Triassic - Jurassic boundary probably occurs at about 695m in this *Zephyr* well. Thus, the highest palynomorph assemblage, from 678-681m (preparation MPZ 669: Fig. 2) is likely to be from rocks of earliest Jurassic (Hettangian) age and those representing the interval 696-723m (preparations MPZ 675, 679 and 683: Fig. 2) are, with those from lower in the succession, of latest Triassic (Rhaetian) age.

### 2. The "Sealab" SLS 17 Borehole

This borehole was drilled on an inlier of Lias situated between the *Zephyr* 87/16-1 and 88/2-1 sites (Fig. 1). It encountered about 60m of grey calcareous mudstones with some thin limestones overlying calcilutites and limestones with thin calcareous mudstone beds which were proved to the terminal depth of 78.9m below seabed. The borehole commenced in early Sinemurian deposits and Hettangian rocks were proven to 49.04m at which

Figure 2. The stratigraphic distribution of palynomorphs in late Triassic and earliest Jurassic deposits in the *Zephyr* 88/2-I and "Sealab" SLS 17 boreholes; the relative abundances of palynomorphs recorded from the "Sealab" borehole are based upon counts of 200 specimens. (*All preparations are held in the Institute of Geological Sciences micropalaeontology collections; those from the "Sealab" borehole are registered in the CSA series and those from the *Zephyr* borehole are registered in the MPZ series).

depth macrofaunal evidence indicated a *liasicus* or *planorbis* Zone (*johnstoni* Subzone) age. The underlying beds are considered to be of earliest Hettangian and possibly late Rhaetian age and may comprise the basal beds of the Lias with, below 60m, the Langport Member of the Lilstock Formation (Penarth Group).

Palynomorph assemblages comprising miospores and organic-walled microplankton, associated with scolecodonts and test-linings of foraminifera, were recovered from eight preparations (CSA 1798-1805) made from core samples taken between depths of 51.5 and 78.9m below seabed (Fig. 2). These assemblages are profuse but poorly preserved and of limited diversity. They are dominated by miospores, principally *Classopollis torosus* (Reissinger) Balme 1957; *Kraeuselisporites reissingeri* (Harris) Morbey 1975 is consistently present but other palynomorphs mostly occur only sporadically and in very small numbers (Fig. 2). The assemblages are comparable with those recovered from above 723m in the nearby *Zephyr* 88/2-1 well (Fig. 2) and with those of late Rhaetian to early Hettangian age obtained from the basal Lias and the underlying Langport Member of the Lilstock Formation (Penarth Group) proved in the Winterborne Kingston borehole (Warrington, 1982). The position of the Triassic-Jurassic boundary is not determined by palynological criteria (see *Zephyr* 88/2-1, above) and the palynomorphs recovered from the SLS 17 samples neither prove nor disprove the presence of rocks of Rhaetian age.

The assemblages do, however, indicate that the borehole penetrated to beds no older than equivalents of the Langport Member and this palynological evidence supports the view that the beds proved below 51.5m comprise lowest Lias and possibly highest Penarth Group deposits and are of early Hettangian and possibly late Rhaetian age.

## Regional summary

The palynomorph successions known from Late Triassic and earliest Jurassic sequences at outcrop in Devon and Dorset and those documented here, from sequences of the same age proved at sites between 100 and 200km south-west of those outcrops, are remarkably similar. Variations in the composition and diversity of palynomorph associations in the outcrop areas correspond closely with facies changes and are regarded as largely' environmentally induced and related to the onset and progress, during the Rhaetian, of a transgression which resulted in the establishment of an open-sea regime throughout much of the British Isles by Hettangian times (Warrington, 1981, 1982). The characters of the palynomorph associations from the *Zephyr* and "Sealab" sites indicate that the sequence of environments in which successive units in the higher part of the Mercia Mudstone Group and the Penarth Group of Devon and Dorset were deposited was also manifest up to 200km farther south-west, in the present offshore region.

No comparable records of Late Triassic and earliest Jurassic palynomorphs are yet available from more westerly offshore areas. An assemblage of miospores of ?Norian - Rhaetian age from the "Sealab" SLS 72 borehole, some 120km west of Lands End (Fig. 1), was summarised by Wilkinson and Halliwell (1980). Late Triassic palynomorphs have also been reported from BNOC well 72/10-1A, situated near the edge of the continental shelf (Fig. 1) about 285km south-west of Lands End (Fisher and Jeans, 1982).

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# The possible influence of storms in the deposition of the Bude Formation (Westphalian), north Cornwall and north Devon

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The depositional environment of the Bude Formation has been a controversial issue for over twenty years (Reading 1963). Although it is generally agreed that deposition took place within a narrow, fresh to brackish basin that straddled much of south-west England (Anderton and others 1979, pp. 160-162 and Fig. 10.8(d)), there are still profound differences of opinion as to the precise environment. The problem stems from the fact that none of the Bude Formation's three constituent facies (black shales, thick sandstones "with few visible structures", and alternations of sandstones, siltstones and shales; Reading 1963) has yielded reliable *positive* evidence of the depth of deposition. *Negative* evidence, particularly the absence of definitive "shallow water" and "coastal plain" features *(ibid.)*, is useful but ambiguous. Amid this climate of uncertainty, two conflicting schools of thought have emerged.

A deltaic environment has been Proposed by a number of authors (Owen 1950; Prentice 1962; King 1967; Edmonds and others 1968; Freshney and others 1972, 1979), some of whom appeal for support to the gross lithological similarity between the Bude Formation and the partially contemporaneous Bideford Formation of north Devon, which is demonstrably deltaic (Raaf and others 1965; Elliott 1976; Edmonds and others I979). A deficiency in this interpretation is that the Bude Formation lacks many of the diagnostic features of the Bideford Formation, including large-scale coarsening-upward cycles, small-scale fining-upward sequences, "medium-scale cross bedding", and "sharply erosive" channels (Reading 1963).

The second body of opinion considers the Bude Formation to be of deeper-water origin, based primarily upon the recognition of turbidites (Ashwin 1957; Lovell 1964) and the absence of unequivocal shallow-water indicators (Reading 1963); some members of this school have suggested that certain bed-thickness trends reflect deposition on a "relatively deep" subaqueous fan (Burne 1969; Melvin 1977). A possible weakness of deeper-water models, however, is the presence of xiphosurid tracks, which may indicate comparatively shallow depths (King 1965; Goldring and Seilacher 1971).

As mentioned above, the distinctive thick sandstones of the Bude Formation have been noted for their lack of visible sedimentary structures. However, during recent field work near Bude, the author observed several examples of a structure resembling hummocky cross-stratification (HCS; Fig. 1), although the effects of compaction and weathering render positive identification difficult.

The term "hummocky cross-stratification,' was introduced by Harms and others (1975) for a primary sedimentary structure with the following essential characteristics: ()) lower bounding surfaces of sets are erosional and commonly slope at angles of less than ten degrees, though dips can reach fifteen degrees; (2) laminae above these erosional set boundaries are parallel to that surface, or nearly so; (3) laminae can systematically thicken laterally in a set, and (4) the dip directions of the erosional set boundaries and of the overlying laminae are scattered *(ibid.)*. The author submits that all of these features are displayed, with varying degrees of clarity, by the thick Bude sandstones. Furthermore, Harms and others (1975, 1982)have reported that HCS is mainly confined to sediments in the coarse-silt to fine-sand range, and that intervals with HCS are generally devoid of both trough and tabular cross-stratification: both of these characteristics are applicable to the Bude Formation.

Although HCS has been neither experimentally produced nor observed forming in modern sedimentary environments, consideration of its facies associations and stratification characteristics has led to widespread agreement that the structure is formed by intense oscillatory flow associated with storm waves (Harms and others 1975, 1982; Dott and Bourgeois 1982).

Assuming that HCS is indeed a product of storm-wave activity, the following important corollaries will ensue if forthcoming field work confirms its presence in the Bude Formation. Firstly, storm-related processes may be invoked as having had a significant influence on the deposition of the Bude Formation. Secondly, depth constraints may be placed upon the depositional
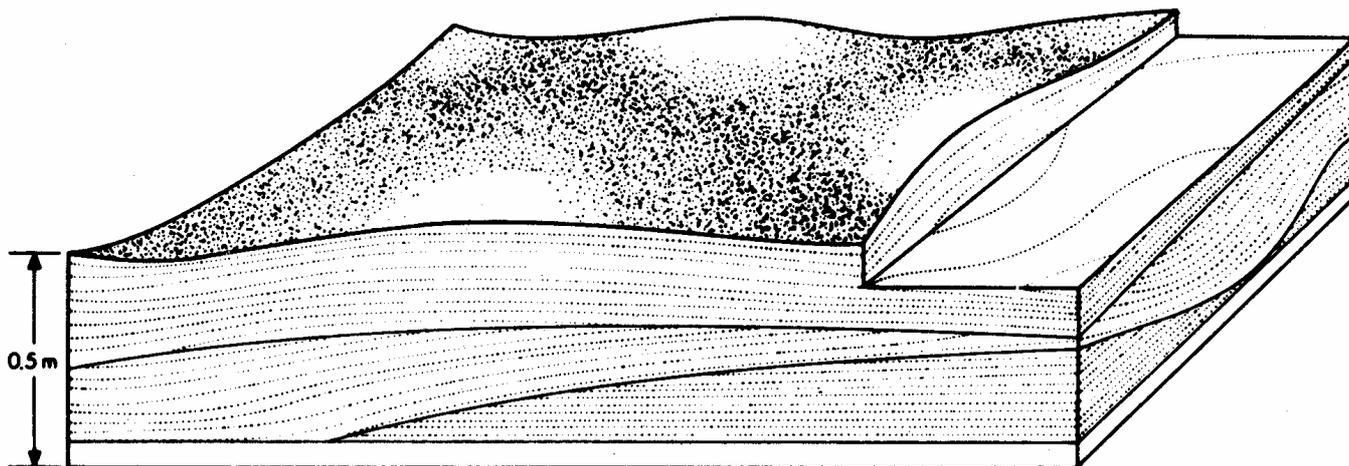
Figure 1. Block diagram showing three-dimensional characteristics of hummocky cross-stratification (from Harms and others 1975).

environment of HCS-bearing intervals, since the structure is though to be formed *above* storm wave base, but preserved only *below* fair-weather wave base (Harms and others 1982). Data from modern continental shelves (Elliott 1978; Johnson 1978; Harms and others 1982) suggest that depths of 200m and 20m, respectively, might provide reasonable approximations of these limits. Finally, the presence of HCS would imply that the Westphalian basin of south-west England was not only of suitable depth, but also sufficiently elongated, parallel to the prevailing palaeowind, for the generation of large waves.

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# Notes on the hydrogeology of the Plymouth Limestone

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## The Plymouth Limestone and its place in the water supply of Plymouth

The Plymouth Limestone is the most westerly of a number of separate Middle Devonian carbonate outcrops occurring in South Devon, (Taylor 1951, Orchard 1978) which although isolated have been considered by Braithwaite (1967) as being of broadly comparable age and lithology. It is recognised however that these similar but separate outcrops of limestone were neither deposited as a continuous sheet nor exactly contemporaneously (Elwood 1982). For the present purposes the Plymouth Limestone is taken as the limestone outcrop centred on Plymouth extending from Cremyll in the west through Cattedown, to Pomphlett Mill Quarry in the east, as shown in figure 1. The outcrop extends for some 10kms from east to west with an average width of l.Skms and covers an area of about 8.15 sq kms. The general elevation of the limestone is much lower than that of slates to the east and west and averages about 30 metres A.O.D,

Prior to the construction of Drake's Leat in 1590 which brought water from the Meavy catchment to Plymouth, the town's water supply was partly from wells sunk into the limestones and slates. The names of some of these wells survive in street names such as Buck Well and Lady Well. These early wells, some of which can still be seen, fell into disuse as piped water became available. The nineteenth century witnessed a resurgence in the provision of water from bore holes including amongst others the Victoria Spa well (2) in Bath Street ( 1841) and the Octagon Brewery well (3) in Martin Street (1896). Flows from these wells were as high as 140,000 thousand gallons per day from bores up to 143 metres deep (Ussher 1907).

There are no records of any systematic examination of the hydrogeology and water availability of the Plymouth Limestone with the exception of a short account by Ingliss in 1877. These notes seek to bring together what is currently known of the hydrogeology of the Plymouth Limestone. Lack of interest in the hydrogeology of the Plymouth Limestone is understandable following-the construction of Burrator Reservoir in 1898 which in combination with direct abstraction from the River

Meavy provides Plymouth with a good supply of water. Prior to these two developments however, the old wells were of intermittent importance in dry summers when the old leat dried up, and in hard winters such as 1881 when it froze (Bracken 1931).

## Lithology and Structure

Braithwaite (1966) and Orchard (1978) have shown that lithologically the Plymouth Limestone is quite variable ranging from unfossiliferous slaty limestones interbedded with shales and tuffs through thin-bedded dark limestone to massive highly fossiliferous limestones. They are believed to have formed as a widespread sheet in shallow marine conditions on a locally uplifted schwelle. Hydrogeologically it is important to note the predominantly dense crystalline nature of the limestone with, as a result, extremely low or nil primary intergranular porosity and permeability. As a consequence with the exception of bedding the hydrogeology of the limestones is almost entirely dominated by secondary features related to the complex structural history of the area.

Taylor (1951) interpreted the structure of the Plymouth Limestone as one large east, west recumbent syncline faulted on its northern margin and reclined to the north, whilst Ussher (1907) and Worth (1891) described a series of smaller fold structures at Cattedown and Richmond Walk. Taylor's interpretation was challenged by Braithwaite (1965) who interpreted the limestones as a single southerly inclined sheet with a gently undulating dip, locally overfolded and thrust along its southern margin. More recent work by Chapman *et al* (in press) suggests the single sheet as proposed by Braithwaite (1965) has been thrust as a unit northwards over Upper Devonian slates. There may be hydrogeological evidence for the thrust proposals of Chapman *et a*/with limestones appearing within and beneath shales in wells at the Bedford Brewery (4) near the City Centre, and the Regent Brewery at Stonehouse (Ussher 1907). The outcrop is traversed by at least four large faults (Taylor 1951,

Orchard 1978) as shown on figure 1 and is cut by two major joint sets both striking between 80° and 100° with dips to the north and south.

The structurally controlled fissure flow system within the limestones has been enlarged by solution to form the cavern systems referred to by Ingliss (1877), Worth (1879) and Ussher (1907). Ewers (1978) has shown that solution passages form from the recharge area to the discharge area. Following Ewers model the concentration of external cavern systems towards the southern margin of the Plymouth Limestone at Mountwise, Stonehouse and the Hoe may indicate recharge from the slates to the north. These cavern systems have probably been extended vertically by fluctuations in sea level in excess of 20m as evidenced by raised beach deposits from the Hoe (Moore 1841) and elsewhere in south-west England (Orme 1960). The limited groundwater level data demonstrates a gradient seawards from north to south in central region and from east to west in the east of the outcrop.

## Water Quality

Typical analyses of uncontaminated limestone waters are shown in table one. Analyses carried out in 1841 on water from the Victoria Spa (2) (Ingliss 1877) clearly showed contamination by sea water and similarly The Octagon Brewery Well (3) was proved saline in 1896 (Ussher 1907). More recent analyses of well water would appear to confirm saline contamination of a number of wells at Stonehouse, Millbay, Cattedown and the Breakwater quarry. Ingliss (1877) reported a series of intertidal springs around the limestone fissure systems. Wells near the shore such as the Lamos Gelatine well (5) at Cattedown can be shown to be altering their water level in sympathy with tidal movements.

The fissure flow regime can have important pollution implications, providing a rapid flow path for polluted surface waters without any of attenuation and filtration mechanisms now associated with intergranular flow mechanisms. This can result in serious bacteriological contamination. A well used by Albany Meat products at Billacombe Road, Plymstock (6), some 58m deep and licensed for an abstraction of 8.6 m.g.y, was grossly polluted in late 1969 by bacteria and other organic matter giving a total suspended solids content of 680mg/1 and a biological oxygen demand of 6.5mg/1. (South West Water Authority Well Record 2/G/Y). The use of the well was subsequently abandoned. This is not a new problem as reports exist of houses in Emma Place (Ingliss 1877) for example, at one time using caverns beneath the area as cesspits and bomb damaged or otherwise leaky sewers must always pose a threat to the quality of water within the limestones.

## The Size of the Resource

Direct recharge to the limestones is about 650 million gallons per year, based upon an average annual rainfall of 95mm and average annual potential evapotranspiration of 576mm. An additional recharge is almost certainly taking place from the Upper Devonian slates which outcrop immediately to the north of the limestone. It is of note that these slates have also provided significant quantities of water during the last Century. Total licensed abstractions from wells and bores sunk in and adjacent to the Plymouth Limestone, have been as high as 214 m.g.y. (32% of the recharge) but are currently only 168 m.g.y. (25% of the recharge). Actual abstractions however are estimated to be only 65 m.g.y. (10% of the recharge) (Table 2).
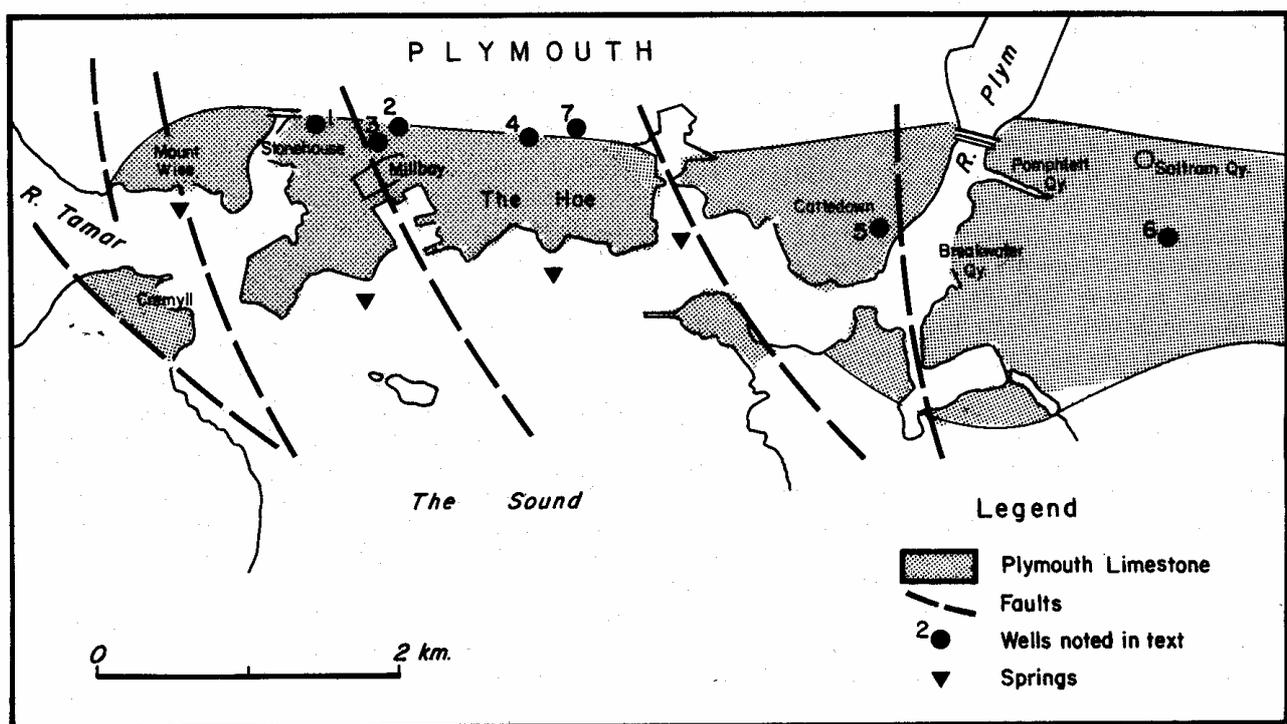


Figure 1. The Plymouth Limestone outcrop showing faulting, springs and wells and locations referred to in the text.

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## Table 2. Licenced Abstraction Current Licences (SWWA)

| Well No. | Well Site | | Licenced Abstraction mg/y | tcm/y |
|---|---|---|---|---|
| 1 | Regent Brewery, Stonehouse | | 10 | 45.4 |
| 5 | APCM, Plymstock Quarry | | 60 | 272.6 |
| 6 | Three wells ........................................... 1 | | 39 | 177.3 |
| | ECC Quarries, | | | |
| | Moorcroft Quarry................................ 2 | | 39 | 177.3 |
| | Billocombe ......................................... 3 | | 20 | 90.92 |
| 8 | Elburton Stud Farm | | 0.073 | 0.332 |
| | Total annual licenced abstraction = | | | 764.012 |

### Past Licences on Record (SWWA)

| Well No. | Well Site | | mg/y | tcm/y |
|---|---|---|---|---|
| 6 | ECC Quarries, Billacombe ............... 4 | | 22.5 | 102.285 |
| 7 | Elburton Vineries | | 0.04 | 0.18 |
| 16 | Albany Meat Co. Ltd., Billacombe | | 8.6 | 39.095 |
| 15 | Spillers Bakery | | 15 | 68.19 |
| | | | | 209.75 |
| | | | | 764.012 |
| | Maximum annual licenced abstraction | = | | 973.762 |

## Table 1. pH levels and ion concentrations of well water samples

### 1. Regent Brewery (Stonehouse) (pH 7.602)

| ion | ppm | epm | % epm(a) | % epm(b) |
|---|---|---|---|---|
| $Ca^{2+}$ | 26.67 | 1.331 | 55.55 | 55.55 |
| $Mg^{2+}$ | 5.46 | 0.449 | 18.74 | 18.74 |
| $Na^+$ | 8.28 | 0.36 | 15.02 ) | 25.7 |
| $K^+$ | 1.0 | 0.256 | 10.68 ) | |
| | | | | |
| $HCO_3^-$ | 475.8 | 7.799 | 70.35 | 71.29 |
| $SO_4^-$ | 74 | 1.54 | 13.89 | 14.08 |
| $Cl^-$ | 56.8 | 1.6 | 14.43 | 14.63 |
| $NO_4^-$ | 9.1 | 0.147 | 1.32 | |

### 2. Prysten House (pH 7.604)

| ion | ppm | epm | % epm(a) | % epm(b) |
|---|---|---|---|---|
| $Ca^{2+}$ | 31.9 | 1.592 | 58.98 | 58.98 |
| $Mg^{2+}$ | 3.66 | 0.301 | 11.15 | 11.15 |
| $Na^+$ | 7.36 | 0.32 | 11.86 ) | 29.87 |
| $K^+$ | 1.9 | 0.486 | 18.01 ) | |
| | | | | |
| $HCO_3^-$ | 436.6 | 7.156 | 71.82 | 72.9 |
| $SO_4^-$ | 74 | 1.54 | 15.46 | 15.69 |
| $Cl^-$ | 39.76 | 1.12 | 11.24 | |
| $NO_4^-$ | 9.1 | 0.147 | 1.47 | |

### 3. Pomphlett Quarry (pH 7.54)

| ion | ppm | epm | % epm(a) | % epm(b) |
|---|---|---|---|---|
| $Ca^{2+}$ | 60.9 | 3.039 | 72.62 | 72.62 |
| $Mg^{2+}$ | 7.77 | 0.639 | 15.27 | 15.27 |
| $Na^+$ | 9.6 | 0.2 | 9.78 ) | 12.01 |
| $K^+$ | 1.2 | 0.307 | 3.73 ) | |
| | | | | |
| $HCO_3^-$ | 433.1 | 7.098 | 75.19 | 76.59 |
| $SO_4^-$ | 59.7 | 1.24 | 13.13 | 13.38 |
| $Cl^-$ | 33.015 | 0.93 | 9.5 | 10.03 |
| $NO_4^-$ | 10.7 | 0.172 | 1.82 | |

### 4. Saltram Quarry (pH 7.733)

| ion | ppm | epm | % epm(a) | % epm(b) |
|---|---|---|---|---|
| $Ca^{2+}$ | 60.9 | 3.039 | 74.83 | 74.83 |
| $Mg^{2+}$ | 10.7 | 0.88 | 21.67 | 21.67 |
| $Na^+$ | 0.92 | 0.04 | 0.98 ) | 3.49 |
| $K^+$ | 0.4 | 0.102 | 2.51 ) | |
| | | | | |
| $HCO_3^-$ | 518.6 | 8.5 | 74.57 | 76.03 |
| $SO_4^-$ | 88.3 | 1.84 | 16.14 | 16.45 |
| $Cl^-$ | 24.82 | 0.84 | 7.36 | 7.51 |
| $NO_4^-$ | 13.5 | 0.218 | 1.912 | |

# Observations on the development of overbank sediments in the Narrator valley, Dartmoor

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## Introduction

The Narrator catchment on south-west Dartmoor (SX 575687) occupies an area of approximately 4km². The upper part of the drainage basin is open moorland. Further downvalley, extensive areas were the scenes of considerable former tin-streaming and farming activity, while the lowest part of the basin was forested by the City of Plymouth Water Department in the 1920's. Recent plantation improvement, involving the trenching of drainage ditches, revealed deposits which indicate changing palaeohydrological conditions in the lower part of the Narrator Brook valley.

## The Sampling Sites

Many trenches soon become overgrown and are subject to caving, thus obscuring and disturbing the deposits. One very new ditch was chosen for this study and sampling was undertaken whilst the exposures were still 'fresh'. Observations in the older ditches showed similar sediments to those of the study trench and it was felt that detailed sampling at four sites along a 50m length of this ditch would be representative. Site 1 is at the downvalley end of the ditch and each other site a further 10m apart.

The general sequence of deposits over the south bank area of the Narrator Brook is made up as follows. A basal peat, termed by the authors 'Narrator Peat Bed'. This is 0.5m to >1m thick and lies directly upon decomposed granite. Above this are silty-clays, sands and gravels totalling 0.75m capped by another organic unit which was named 'Rough Tor Peat'. Above are small lenses of fine gravels and coarse sands interdigitating laterally with clays and passing upwards into modem soil and made ground.

*Site 1:* The most distinctive feature of this site is an extremely well preserved 'flame' structure. The silty clays that compromise the main body of this feature rise upward into an overlying gravel and sand formation. The top of the 'flame' appears truncated by a 250mm mottled brown and grey-brown silty clay unit into which two fining upward gravel to sand sequences pass laterally. The section is capped by 70mm of dark brown silty clay with peaty laminations and a generally high peat content. This latter unit could be a disturbed horizon where silty clay has been mixed with Rough Tor Peat.

*Site 2:* Here is a mixture of silty clays and coarser higher energy deposits. A 50mm grey silty clay unit is present above the Narrator Peat. This passes upward into a strongly convoluted lobate feature composed of light grey-brown silty clay with peat inclusions. This horizon has been forced upward into an overlying 100mm coarse sand and medium gravel unit. Strongly mottled silty clays with peaty inclusions overlie these coarse members before passing upward into a disturbed unit which includes a thin lens of Rough Tor Peat.

*Site 3:* The upper peat is absent from this sample site. In detail, the Narrator Peat is overlain by a total of 350mm of predominantly coarse deposits truncated by 150mm of grey-brown and brown silty clays. Abruptly overlying this are four repeated fining upward units of well sorted fine gravel, coarse sand and medium to fine sand. Fibrous organic material is present throughout the 220mm thickness of this sequence. At the top of the section is 80mm of mottled brown silty clay with peaty inclusions, presumably derived from erosion and fragmentation of the Rough Tor Peat.

*Site 4:* Between the Narrator Peat and the Rough Tot Peat at this site are 0.5m of silty clays with irregular laminations, peat inclusions and colour variations. Here the Rough Tor Peat is 100mm to 120mm thick.

## Sedimentary Structures

The origin of the 'flame' structure of Site 1 was first attributed to gravity loading from above, modified by plastic flow and current deformation. The planar character of stratification below and the sharp truncation at the top of the contorted zone is of the type described by Dott and Howard (1962). However, the abundance of organic material below this 'diapir' would seem to favour the model suggested by Coleman and Gagliano (1965). They suggested that where sandy sediments accumulate rapidly over deposits rich in organic matter, gases from the decomposing organic material rise or heave upward through the overlying sediments, producing distortion.

Convulate laminations are relatively common in the study trench and occur in the silty clay units. Many convolutions appear as slightly asymmetrical contortions

of the 'drag' type discussed by Friend (1965). The work by McKee, Crosby and Berryhill (1967) on flood deposits in Colorado suggested that convolute structures were developed during a late stage of flooding when current velocities had slowed and sediment was in the condition of quicksand. However, Sanders (1960) attributed the formation of such structures to shearing effects set up by high velocity currents. With the convolute structures seen in the Narrator sections, it is not possible to attribute their origin definitively to any one specific hypothesis.

Fining upward sequences occur as distinctive features of the middle sections of the study ditch. As Allen (1964, 1965a and b) commented, such fining upwards is very common especially in channel deposits. Sites 2 and 3 show the structures most clearly with repeated sequences indicating deposition from successive flood events moving through an overspill or crevasse-splay channel. The model described by Allen (1965a) depicting localised tongues of sediment, sinuous in plan, deposited over the outer margins of floodbasins from crevasse channels that tapped active streams in flood, seems to fit the observed nature and distribution of sediments and structures in the area.

## The Nature of Fluvial Deposition

The occurrence of peats, silts and clays in combination with sands and gravels, have been noted as facets of the fluvial environment (Allen, 1964 and 1965a). In general, silt and clay may be said to have been transported as washload, sand and gravel as bedload. In view of the lack of features indicative of sub-aerial exposure in the study sections and the thickness of sediment which constitutes each unit, it seems unlikely that the fine units were deposited on a floodplain proper. These units form a single, though variable, textural group and are interpreted as being floodbasin deposits. There are few homogeneous silt and clay units, no internal stratification or lamination and bedding defined by marked lithological contrasts is not generally apparent. It is likely, therefore, that the water was periodically stirred up by overbank flows, which also supplied a fresh input of sediment (Daley, 1973).

The movement of gravel as bedload clearly requires the existence of a channel for the maintenance of relatively high shear velocities: Crevasse-splay deposits are noted by Allen (1965a) as consisting of moderately well sorted sediment units which are similar in grade to the bedload and channel deposits of the main stream. They range from 10mm to over 1m in thickness, often contain drifted organic material and may be cut into or cover existing vegetation. The coarse units exposed in the study trench have an average sorting coefficient of 1.4 and range from moderately sorted to poorly sorted deposits. Although they are not completely analgous to crevasse-splay deposits as defined by Allen (1965a), it is proposed that they be classified as such. The fining upward sequences

are interpreted as being formed during periods of waning flow when the coarsest sediments were swept into the floodbasin and deposited.

## Dating

As the fluvial sediments overlie weathered granite and thin out against solifluucted slope deposits, they must post-date the Pleistocene cold phases. Samples of the peats were analysed at Harwell and given radio-carbon dates of 630BP and 4040BP + 60. Thus the floodbasin and crevasse-splay deposits originated between these dates. Simmons (1964) has noted that tree clearances took place on Dartmoor during the Bronze Age. The authors suggest that the deposits described be attributed to hydrological changes brought about by forest clearance. The activities of tin streamers from the 12th to 19th centuries (Greeves 1969, Cook et al., 1973), with their drainage diversions, may have brought deposition in the floodbasin to an end.

## Conclusions

The deposits exposed in a drainage ditch in the Lower Narrator Valley are considered to be of fluvial origin, resulting from periods of high flow when the nearby Narrator Brook overtopped its banks, creating a 'floodbasin'. Three types of deposit are identified: namely peat, fine (silt-clay) units and coarse (sand-gravel) units respectively. The sand and gravel units exhibit striking fining upwards cycles and are considered to be the result of major discharge events which swept heavily charged floodwaters into the flood-basin, carving channels through it and allowing bed load from the main channel to be entrained and eventually deposited when the flows declined. The silt and clay units are thought to be the result of periodic overbank sedimentation which simply inundated the floodbasin and from which the silts and clays eventually settled. These coarse and fine units are thus termed crevasse-splay and floodbasin deposits respectively: The peat units are considered to have been formed during periods when vegetation dominated the surface of the floodbasin.

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# Bertrandite from Hingston Down Quarry, Calstock, Cornwall

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Hingston Down Quarry, 2km W of Gunnislake, is a working quarry in the small granite intrusion of Hingston Down, situated midway between the Bodmin Moor and Dartmoor plutons. The fine-grained biotite-granite of Hingston Down is traversed by numerous nearly vertical ENE-trending veins and mineralized joints containing various minerals including chalcopyrite, arsenopyrite, pyrite, sphalerite, tourmaline, molybdenite, fluorite, wolframite and scheelite.

In June 1978, crystals of the basic beryllium silicate, bertrandite, were found here (SX 410717) in a block that had recently been detached from the lowest of the three working faces; three other specimens have since been collected by my colleague Mr B.E. Brett. The bertrandite forms equant tabular crystals up to 7mm across showing a pearly lustre on (001), and intergrown groups, in cavities in greisenised granite containing schorl, with pale green fluorite, chlorite, chalcopyrite and pyrite. On one specimen, from an open joint-fissure in the middle level, dozens of equant bertrandite crystals about lmm across form clusters of intergrown crystals, and an epimorph after fluorite, associated with fluorite and quartz crystals. The late Richard W. Barstow presented two bertrandite specimens from Hingston Down Quarry to the British Museum (Natural History) in July 1980, the associated minerals being orthoclase and fluorite, and arsenopyrite.

Most of the bertrandite crystals from Hingston Down Quarry are six-sided in outline owing to equal development of m(110) and b(010), and resemble in habit and mode of occurrence those found in 1904 at the Cheesewring Quarry, Linkinhorne (Bowman, 1911). Some of the smaller crystals are, however, rectangular in outline, and twinned on e(011), the $60^\circ$ re-entrant angle being formed by the faces of c(001). The smaller crystal has often grown from near the middle of one side of a thicker, wider individual, as Russell (1913) observed in the bertrandite twins from Gold-diggings Quarry, St Cleer.

Beryllium minerals are rare in south-west England. In addition to the first recorded occurrence in Britain (Bowman, 1911), and the four localities described by Russell (1913), bertrandite has been observed in crystal-lined cavities in complex pegmatites at Trolvis Quarry, Stithians (Hosking, 1954, p.280). At all these localities the bertrandite occurs as a primary phase in fissures in granite and is of late-stage hydrothermal origin.

The identity of the bertrandite was confirmed by an X-ray- powder diffraction photograph. The positions of most of the d-spacings agree with published data (P.D.F. card 12-452), but the weakness of the d040,220and d060 reflections compared with stated relative intensities; and absence of the d 350,080 reflection, may have resulted from preferred orientation of cleaved fragments. It also gave an infrared spectrum closely comparable with published data for bertrandite (Farmer, 1974, p.378). The presence of Be was confirmed, after chemical separation, by the formulation in alkaline solution of a cornflower blue lake with quinalizarin.

Spectrographic analyses by Butler (1953, p.160) of samples from five of the Cornish granite masses suggests that the Hingston Down granite may be enriched in Be, his sample (from "W. Chilsworthy") containing 25ppm Be compared with 10-15ppm in other Cornish granites. Determinations of Be in two samples from Hingston Down and one from the nearby, but separate, Gunnislake mass (Ward, 1971) carried out by Mr S.J. Adams using atomic absorption, proved the fine-grained megacryst-poor biotite-granite of Hingston Down Quarry to contain 6ppm, whilst a leucogranite, which forms veins and tongues in the former, contained 33ppm Be. A coarse-grained poorly megacrystic lithionite granite from Pearson's Quarry, Gunnislake, contained 15ppm Be. At Hingston Down, Be is thus concentrated in the residual magma, and the frequent association of fluorite with bertrandite in greisenised joints in the granite here, and in the other Cornish occurrences, suggests migration of Be in aqueous F-bearing solutions, perhaps as mobile complex fluoroberyllates in the hydrothermal stage. Reaction with wall rocks may then have caused a change of pH and hydrolysis of unstable fluoroberyllate ions (Beus, 1966), resulting in formation of bertrandite and fluorite.

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# Deformation of weathered profiles, below head, at Constantine Bay, North Cornwall (Abstract)

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Deformation of weathered bedrock, involving the downhill deflection of inclined strata, is frequently found below head throughout south-west England. Recent studies (Mottershead, 1971; Green and Eden, 1973) have considered these structures and regarded them as being the result of downslope drag by Pleistocene soliflucting head. The present investigation concerns a hitherto unstudied site of deformation structures restricted to weathered Middle Devonian grey slates at Constantine Bay, North Cornwall (SX 744856). The aim of this present structural study is to investigate the mechanisms of these sub-surface deformations and their relationships, if any, to soliflucted head deposits.

Within a 100m long coastal cliff section, south of the entrance to Constantine Bay an upward sequence may be defined of 1. Grey slates, 2. 0.15-1.2m thick frost-riven weathered slates, 3. 0.2-1.0m thick silty-clay head, 4. Recent wind-blown sands. The head deposit can be differentiated into a lower grey soil unit with slate lithoretics, passing up into weathered brown soil.

Long-axis stone fabrics of the head have distributions that are co-planar, with the head slope. Maximum concentrations of stone long axes are parallel to the slope dip, implying downhill solifluction of the head unit.

The weathered profile is readily distinguished from the underlying grey slates by a contrast in weathering grade and cleavage-parallel, fracture spacing. A sharp change in fracture spacing marks the base of the weathered profile. The parallel attitude of the weathered profile base to the local head slope suggests that its formation was influenced by the pre-existing hill-slope form. The widespread development of ice-wedge casts, penetrating the full extent of the weathered profile, provides convincing evidence of periglacial activity on these slopes. The termination of ice-wedge casts above the weathered profile base may indicate that the latter was the permafrost table and the overlying ground the seasonally-thawed active layer.

Two distinct deformation modes are restricted to the weathered slate profile:

(a) Bedding folds. Where bedding, inclined to cleavage. can be traced through the weathered profile it is seen to be deflected.

(b) Cleavage folds. Kink and conjugate style folds of cleavage planes

In both structural modes a deformation mechanism involving slip along cleavage-parallel fractures has been recognised. The passive displacement of bedding by cleavage slip has been analysed using a simple-shear strain model. The model provides both a quantitative estimate of shear strain variation and an explanation for the geometrical variations shown by cleavage fold elements. Two types of shear strain variation have been recognised. In type 1 the shear strain intensity increases progressively upwards through the profile. In type 2 the shear strain shows a similar increase in the lower part of the profile, whereas in the upper part type 2 shear strain decreases, such that only a small deformation is recorded at the top. Type 2 deformation indicates that the upper sections of these weathered profiles have behaved as rigid plugs with little or no internal deformation. The variation in shear strain intensity is closely matched by the deformation intensity registered by cleavage folds. Thus with increasing amounts of deformation, cleavage folds change from open symmetric, upright folds into tight, asymmetric forms with the axial surface progressively rotated in the direction of shear. A spatial analysis of the shear strain profiles suggest lobe-like displacement surfaces whose median lines parallel the cleavage dip direction.

The restriction of sub-surface deformation and ice-wedge casts to slate weathered profiles, appears to support the view that deformation arises from cryogenic movements within a former permafrost active layer. That solifluction movements took place is indicated by head stone fabrics and the downslope stripping of regolith from the southern higher ground. The form of some deformation gradients implies the existence of a rigid crust which has undergone passive transportation during deformation. The precise cause of these movements remains unsolved. It is believed that high pore fluid pressures may have played an important role in a former periglacial deformation of the slate regolith. This view is supported by the existence of confined aquifers within the active layer of present tundra environments (Washburn 1979). The confinement of such an aquifer may well reflect either spring-summer thaw or autumn-winter freezing conditions. In either instance aquifer confinement by a rigid surficial plug seems plausible, if by the action of a downward moving freezing front or by the formation of an impervious weathered crust.

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# Distribution of radioactive isotopes in relation to groundwater movement in south-east Devon (Abstract)

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South-east Devon is an area in which the geological formations range in age from Devonian to Recent, but most ground is composed of the generally easterly dipping New Red Sandstone Series (Stephanian - late Triassic). Only in the extreme west of the area, and in a belt stretching east to Exeter, are highly deformed pre-New Red Sandstone formations found. Easterly dipping Jurassic rocks which conformably overlie the New Red Sandstone, are confined to the eastern margin of the area, beyond Axminster. Cretaceous and Tertiary rocks generally lie more or less horizontally in an overstepping relationship with the older rocks, but in the Bovey Basin they have been downwarped/faulted along the line of the Sticklepath Fault. Topographically the Cretaceous and Tertiary strata form the high ground of the East Devon plateau and the Haldon Hills, but the plateau is extremely dissected by southward flowing rivers so that these rocks now form the hill cappings.

Heat flow in south-east Devon is above average for the United Kingdom as a whole, but values fall rapidly to the north, and in the Bath-Bristol area they are well below average, although at Bath and Bristol hot springs occur. The distribution of radioactive isotopes suggests that modern hydrothermal circulation is taking place within south-east Devon.

Radon concentrations in the stream waters of south-east Devon show the presence of a zone of anomalously high values extending from the mouth of the Exe Estuary, north-north-westwards to Kennford, and on to near Crediton. This is interpreted as being caused by groundwater discharge along a fracture zone related to the system of transcurrent faults found with this trend in south-west England. Apart from a coincidence of high values at Kennford and close to Exmouth, the distribution of high uranium concentrations in stream water is different from the radon occurrences. Zones of high stream water uranium which occur north of Littleham Cove and between Sidmouth and Honiton, are interpreted as being caused by the presence of secondary uranium-bearing concretionary nodules within the New Red Sandstone. The high values at Kennford are thought to result from secondary uranium precipitation along fractures. Gamma-ray spectrometry of soils in south-east Devon shows a pattern of uranium and potassium distribution in which high values are associated with the mudstone and breccia formations of the New Red Sandstone, but there are also distinct zones of high values related to the north-northwest-south-southeast trending

feature which passes through Kennford, and east-west features which pass through Feniton and the Exmouth-Sidford area. The interpretation of the east-west lines as fracture zones, along which uranium and/or potassium enrichment has occurred, follows from the characteristics of the north- northwest - south - southeast trending zone. The Feniton zone is considered to be an expression of the faulted southern margin of the Crediton Trough, while the Exmouth - Sidford area probably represents the combined effects of two adjacent fractures which are the westerly extension of the Abbotsbury Fault system. Thorium contents of soils in south-east Devon generally follow the lithological character of the underlying rocks.

In the Teign Valley, a uranium-rich, thorium-rich, north-south trending zone occurs, which is considered to be a pitchblende - bearing cross-course.

Apart from the indications given by the stream water radon distribution, modern movement of groundwater in south-east Devon is also shown by the disequilibrium in the uranium decay series. A Disequilibrium Factor has been defined in which the gain of recent uranium shows as positive values while recent loss of uranium produces negative values. Although negative values dominate the distribution of south-east Devon, zones of positive values coincide with the east-west line passing through Feniton, with the Exmouth-Sidford feature and with the north-northwest-south-southeast line through Kennford. These positive values indicate modern enrichment of uranium and upward movement of groundwater.

The fracture zones are considered to be reactivated basement structures, with movement having occurred throughout the Mesozoic and Tertiary. Episodes of mineralisation were brought about by the movement of mineralising fluids along these fractures from the basement of Devonian and Carboniferous rocks, and charging of the New Red Sandstone aquifers. When these fluids discharged through the adjacent mudstone horizons, precipitation took place at sites of organic debris or permeability contrast. The driving force for this groundwater movement appears to be developed over a wide area and is not related to local topographic head. A geothermal head is therefore invoked to account for the observed features. The vigour of the hydrothermal circulation has varied considerably in the past, with tectonism affecting the fracture permeability, and mantle activity the thermal input. The system operating today, however, is a low-energy expression of more vigorous circulation in the past.

# The marine Lower Devonian of the Plymouth area (Abstract)

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The east side of Plymouth Sound displays extensive exposures of the Lower Devonian Meadfoot Group (Harwood, 1976) comprising the Meadfoot facies (Evans, 1981) and the Staddon facies (including the type area of the old Staddon Grits). The onset of deposition of the Staddon facies was not synchronous across the whole of S.W. England but at Plymouth Sound the Meadfoot/ Staddon boundary corresponds approximately to the Siegenian/Emsian boundary on the evidence of the brachiopod faunas.

Brachiopod faunas from the Meadfoot Group suggest a Benthic Assemblage 2 position (Boucot, 1975) for the Staddon facies (i.e. low intertidal) and a B.A. 3-5 position for the Meadfoot facies (i.e. subtidal). This broadly agrees with the lithological and sedimentological evidence from the Plymouth area with subtidal (outer shelf-Goldring and Langenstrassen, 1979) conditions in the Meadfoot facies but with evidence of shallowing in the Staddon facies (inner shelf).

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# The Bristol Channel Landmass (Abstract)

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In Lower and Middle Devonian times, short lived and geographically restricted source areas were active in the region of the present Bristol Channel. Evidence for this local source, shedding detritus northwards, comes from the presence of quartzite, rhyolite, lithic greywacke and phyllite clasts in the southerly-derived, ?M. Devonian Ridgeway Conglomerate of S.W. Dyfed (Williams 1971), and the presence of a restricted accumulation of conglomerate containing coarse sub-angular clasts of porphyries, tuffs, lithic sandstones and quartzites in the Lower Devonian Llanishen Conglomerate near Cardiff (Allen 1975). A southerly transport of exotic debris from a source in the Bristol Channel area is deduced for the M. Devonian Rawns Formation (Tunbridge 1980) of North Devon which contains angular clasts of porphyry, tuffs, quartzite, acid lava and lithic sandstone. Geophysical studies (Brooks and Thompson 1973, Brooks and Al-Saadi 1977) give good evidence for a fault bounded region of ? L. Palaeozoic or ? Precambrian rocks in the northern part of the Bristol Channel. It is considered that fault movement, perhaps coupled to late-Caledonian activity, caused short-lived and localised uplift of this source, shedding material north and south to produce only local accumulations of conglomerate. This source did not feed the main clastic wedge of the Hangman Sandstone Group, which petrological studies show to be formed from re-worked lower Old Red Sandstone from S. Wales.

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# Clastic sediments within and beneath the Meneage Formation, south Cornwall (Abstract)

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The Devonian turbidite sequence exposed in south Cornwall coarsens upwards and to the south, with up to 1 km of olistostromes, rudites and volcanics (the Meneage Formation) developed in the youngest part of the sequence in southernmost Cornwall.

The greywackes are quartzose with approximately equal proportions of. feldspar (plagioclase + minor K-feldspar) and lithic fragments. The latter include a variety of fine grained sedimentary, deformed meta-sedimentary and igneous rock types.

Blocks of Ordovician quartzite up to several hundreds of metres in length occur in the olistostromic deposits. Rudites and pebbly arenites interbedded with the olistostromes include a 200m thick sequence of monomict types (in ascending order: mica schist, amphibolite, deformed garnetiferous granite and deformed arenite rudites) together with polymict rudites (up to 100m thick) incorporating a very wide range of clast lithologies. Clasts range up to lm in diameter and from very well rounded angular. Sorting is poor and the deposits are usually massive. The monomict rudites are laterally persistent and interpreted as sheet debris flow deposits derived from compositionally restricted sources in a complex source terrain. The polymict rudites thin rapidly laterally and are locally capped with upward thinning and fining sequences. These are thought to represent channel deposits.

Lithic detritus in the greywackes and rudites is closely comparable, suggesting derivation from similar source rocks. No direct evidence of sediment transport directions has been recorded in south Cornwall. The quartzite olistoliths have been compared very closely with quartzites now exposed in NW France, where the underlying strata incorporate many of the other lithologies seen as detritus in the south Cornish greywackes and rudites. Hence a northward extension of basement and Lower Palaeozoic cover as seen in NW France is suggested as the most likely source. The only possible relic is the Eddystone Reef from which deformed garnetiferous granite and mica schists have been recorded.

# Tear faulting in south-east Somerset (Abstract)

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To the west of Yeovil (G.R.478173) is an extensive area of old quarries which were once worked for the Ham Hill Stone (Toarcian). Many of the quarry faces show horizontal or near horizontal slickensides; of twenty nine bearings of slickensided faces twelve are approximately NNW-SSE and eight are NE-SW. The amount of movement of individual faces is hard to assess but it is probably no more than a few centimetres. The NNW-SSE faces are mainly dextral in the sense of movement whilst the NE-SW set tend to be sinistral. An exposure near the old lime kiln (G.R.480165) shows movement in four directions together with brecciation. Tension fractures set at $45^\circ$ to the slickensides show scalenohedron calcite crystals.

The directions and sense of movements are similar to those associated with Tertiary tear faulting in Devon and Cornwall. Horizontal slickensides have been observed 12km to the east and 8km to the west of Ham Hill but not in any exposures of the Blue Lias to the north. Drummond (1970) has speculated that the N-S Parrett fault lkm to the west of Ham Hill has a dextral displacement. Further to the north-west lies the Watchet-Cothelstone NW-SE dextral fault which has been traced as far as Hatch Beauchamp (Whittaker 1972).

It is possible that the features described above at Ham Hill and in the Parrett Valley reflect underlying control by basement structures. The Ham Hill stone is a trough cross-bedded bioclastic limestone with a lag conglomerate at the base; its long, narrow N-S outcrop is noteworthy.

Furthermore it lies on the NW end of the Mid-Dorset Swell (Drummond 1970) in an area with marked thinning of the Junction Bed and the Inferior Oolite. These are aspects which merit further study.

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# British Triassic palaeontology: supplement 7

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Since the submission of the writer's previous supplement *(Proc. Ussher Soc., 5,* 394, 1982) to his paper on British Triassic palaeontology, the following works dealing with or including aspects of that subject have appeared:

Ambrose, K. 1982. *Geological notes and local details for 1:10 000 sheet SP 17 SW (Tamworth in Arden).* Institute of Geological Sciences, Keyworth, 19pp.

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