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Research into the geology and geomorphology of south-west England

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Edited by G.M Power

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Papers

Notes and Abstracts

Invited Address to the Annual Conference of the Ussher Society, January 1986

Cornubian Quarter-century: Advances in the geology of south-west England, 1960-1985.

D.L. DINELEY

Dineley, D.L. 1986. Cornubian Quarter-century: Advances in the geology of south-west England, 1960-1985. *Proceedings of the Ussher Society 6,* 275-290.

Amongst the numerous disciplines within the earth sciences that have made notable advances in south-west England during the last 25 years palaeontology, sedimentology, stratigraphy and structural studies have added most significantly to our understanding of the geological evolution of the region. Geochemistry and geophysics have greatly aided this work and have also furthered our understanding of the economic geology. Palaeontological advances have largely been in Palaeozoic biostratigraphy, especially by use of conodonts. Stratigraphy has become greatly refined and the origins of formations better understood in the context of an active Cornubian branch of the Rhenohercynian basin and of post-orogenic (New Red Sandstone and Mesozoic) basin-infilling. Cornubian terrain is now regarded as structurally dominated by "thin-skinned', tectonics in south and west Devon and throughout Cornwall, with numerous flat and thin nappes moving northwards while Devonian and Carboniferous sedimentation took place. In north Devon the structures are steeper and simple. Stratigraphic and structural continuity to depths greater than a few kilometres is doubtful. As in mainland Europe, no simple plate tectonic model of Variscan orogeny is indicated here but strong and persistent dextral fault movement may be significant support for a model involving dextral transcurrent fault movement.

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Introduction

The advances of geological research in its broadest sense in Britain has been nowhere more marked or significant than in the south-west of England. In this the Ussher Society has played a central role as a focus for the presentation and discussion of results. Founded in 1960, it has prospered not only as a forum for meetings but also in its *Proceedings,* which have now become internationally recognised as a publication of merit. The membership of the Society has never greatly exceeded 250 but it has included a large number of young researchers in the universities and polytechnics, together with senior academics, industrial, amateur and government earth scientists. When the Regional Office of the Geological Survey was opened in Exeter in 1968 a vigorously pursued programme of remapping the "One Inch" sheets of the south-west was soon to add a welcome impetus to research into the Cornubian terrain.

In the early days of the Ussher Society it was already clear that research topics would fall into place within one or other of the following general themes:-

1. The nature, origin and history of the basins in which the Devonian and Carboniferous rocks were deposited.
2. The effects, causes and timing of earth-movements that gave rise to the structures within the Palaeozoic rocks, and the relationships of these structures to the "Variscan Front".
3. The origins, modes and timing of emplacement of the Cornubian granites and associated igneous rocks.
4. Post-orogenic basin formation and basin-infilling, continental and marine.
5. Post-Cretaceous uplift and diastrophism, and the stripping of the Mesozoic cover.

Important advances in both technology and methodology applied to the earth sciences during these last twenty five years have included rapid chemical analyses (X-ray etc.), geophysical data collecting, computing, scanning and other forms of electron microscopy and remote sensing. The studies of European earth scientists in other parts of the Variscides and in similar terrains elsewhere have provided models and ideas for comparative studies. All of these contributed to several main fields of achievement:-

1. Characterisation of igneous bodies in terms of geochemistry, mineralogy, age and spatial relationships, origins and history.
2. Parameters for the identification of specific sedimentary bodies (facies analysis) and the means of detailed correlation of strata by means of fossils (especially micro-fossils).
3. The recognition of sequential changes during diagenesis and metamorphism.
4. Means of determining stress relationships: plastic flow and brittle fracture structures, leading to structural models explaining complex outcrop

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patterns and relationships in Cornubian terrain.
5. Means of modelling deeper crustal structures.
6. Development of overall tectonic models to explain (a) evolution of the Cornubian basin(s) and orogen; (b) post-orogenic movements and changes of level consistent with models developed for adjacent areas (plate tectonics).

Several publications indicate the progress these studies have made over the years; notable are *British Regional Geology: South-West England (Edmonds et al. 1975), The Geology of Devon* (Durrance and Laming, eds., 1982), the Special Reports of the Geological Society of London on Stratigraphic Correlation (Nos. 5, 7, 8, 9, 10), *The Variscan Fold Belt in the British Isles* (Hancock, ed., 1983), *Variscan Tectonics of the North Atlantic Region* (Hutton and Sanderson, eds., 1984), and sheet memoirs of the Geological Survey. Finally, and appropriately in January 1986, the Geological Society of London published the thematic set of papers on Variscan Structures of south-west England and related areas, a record of the meeting held at Plymouth Polytechnic, in 1984.

A brief survey such as this is cannot cover all the fields of earth science in the southwestern peninsula. Conspicuous by their absence are references to geomorphology and Quaternary research, economic geology (see Nicholas 1980), and marine geology. Only the briefest mention is made of the Mesozoic rocks that belong to the basins adjacent to the old "Cornubian landmass". Other deplorable omissions no doubt will be obvious to one specialist or another - for these the author apologises and pleads lack of space, or ignorance, rather than lack of interest.

There is no sign that earth sciences have in south-west England reached the point where all seems clear or uncontroversial. The picture grows more complex, and there is still the periodic need to re-examine and question what was accepted in the past. Data that may seem insignificant today may be critical tomorrow. When the Ussher Society celebrates its fiftieth anniversary there will, no doubt, be different explanations for some of the phenomena we regard as unquestionable. Perhaps by then a more comprehensive plate tectonic explanation of Variscan geology in western Europe will be in vogue. In view of the progress made over twenty five years that has given today's plate tectonic models, we have every reason to be hopeful.

Palaeontology

While relatively little has been added to Devonian and Carboniferous palaeoecology or palaeobiology from recent discoveries of fossils in the south-west, biostratigraphy has been greatly improved. Stratigraphic and structural revisions have made headway because of the more precise dating and biostratigraphic correlation now possible. This holds for both Palaeozoic and Mesozoic rocks.

Conodonts and ammonoids have proved invaluable in the Palaeozoic, with ostracods and palynomorphs also locally very important. Prior to about 1950 records of ammonoids in south-west England were few but since then House and others have greatly extended the record, recognising many of the European. and especially the German ammonoid zones here (House 1963; House and Butcher 1962, 1973; Prentice 1960; Selwood 1960).

One of the more significant discoveries has been that of the condensed sequence of ammonoid zones at Chudleigh (House and Butcher, 1973). It led to a comparison of that succession to those known on the German *schwellen.*

In recent years Carboniferous ammonoids have been recovered from many localities, especially Dinantian (see George *et al.,* 1976). In his review of the Upper Carboniferous of the region, Ramsbottom (1970) noted the affinities of the Namurian faunas with those of North Spain and of the Westphalian A faunas with those of eastern North America.

Pioneer work on European conodont biostratigraphy began soon after the second world war. Both the Devonian and the Carboniferous systems were sub-divided on the basis of conodont zones, and the conodont successions equated to the ammonoid sequence. Many of the long-known Devonian formations and fossil localities in south-west England have now yielded conodonts of biostratigraphic value. A summary account of many of these is given in Austin and Armstrong (1985), and the presence of virtually all the zones known from the Middle Devonian to lowest Namurian of Germany is demonstrated. Many conodont faunas have been retrieved from limestones and shales from which no identifiable macrofossils are known. Many localities yield long lists of taxa, but a high proportion of the fossils are broken or eroded because of redeposition. A start is made on conodont biofacies analysis.

In south Cornwall pre-Devonian conodont faunas (Barnes *in:* Austin and Armstrong 1985) have been recovered from rocks of a disputed origin (see below p. 278).

Ostracod faunas, also used in German Upper Devonian biostratigraphy (Rabien 1954), were studied successfully for the same purpose by Gooday (1974) while Selwood *et al.* (1984) figure both ostracods and well-preserved conodonts from the Upper-Devonian of the Newton Abbot area.

Work on Devonian benthos in the carbonate facies of South Devon has revealed much new data on rugosa, stromatoporoids and tabulates. Scrutton and his co-workers (1977, 1978; Goodger *et al.* 1984) describe ecological sequences in the coral-stromatoporoid communities and "reefs" of Torbay.

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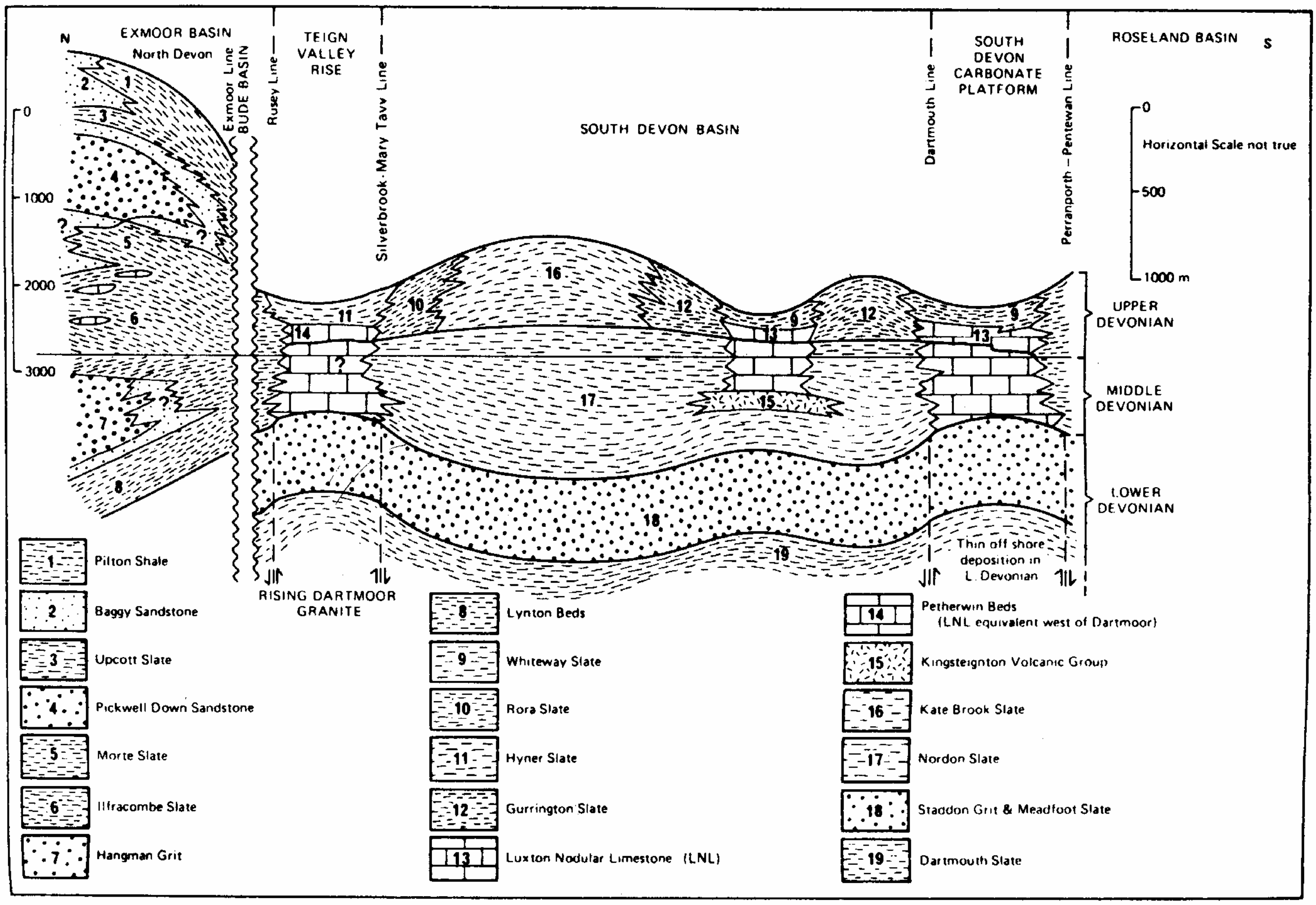

Figure 1. Observed Devonian stratigraphic relationships across Devon (from Durrance and Laming, *The Geology of Devon,* with permission). The postulated Bude Basin Devonian lies beneath the autocthonous-parautocthonous Carboniferous.

Other marine benthos has been found in the Lower Devonian (Evans 1981) and is largely a patchily distributed brachiopod-mollusc assemblage. In north Devon and the Quantock Hills new corals and other taxa have been recorded from the Middle-Upper Devonian (Webby 1965a, b, 1966 a, b; Holwill 1962, 1964) trilobites and brachiopods new to Britain have been found by Goldring (1955, 1971) whose work on the ichnofaunas of the Baggy Beds and Pilton Beds indicates an abundance of both infaunal and epifaunal animals.

Vertebrates in the Gedinnian Dartmouth Beds of the Bigbury Bay, Plymouth and Watergate Bay areas include osteostraci, pteraspids and acanthodians (Dineley 1966). Traquairaspid fragments have also been discovered recently in the Wembury Formation and this may support Hobson's (1976) view that it is a unit near the base of the Dartmouth Beds rather than the top.

In the New Red Sandstone the palynomorphs are increasingly regarded as good stratigraphic indices (see Warrington's bibliography, 1971 onwards). Large trace fossils in the Tot Bay Breccias are now perhaps best interpreted as the burrows of small vertebrates (Ridgeway 1974; Pollard 1976): their like is not apparently known elsewhere in Britain.

Palynomorphs in the Mylor Slates at Mount Wellington in Cornwall indicate a Famennian age (Turner *et al.* 1979) and derived late Permian forms have been recovered from the Mercia Mudstone Formation at St. Audries Bay (Warrington 1979). Miospores from the New Red Sandstone in South Devon include derived Devonian and Carboniferous forms (Warrington 1971).

Work on the Newton Abbot Sheet (Selwood *et al.* 1984) produced a large and impressive fauna of molluscs, bryozoa, corals and other forms from the Upper Greensand of the Haldon Hills area and has enabled a more precise correlation to be made with beds in east Devon.

Stratigraphy and Sedimentology

An understanding of the detailed stratigraphy of the Palaeozoic rocks of south-west England has been hindered by lateral facies variations, by structural complexities and by the irregular distribution and poor

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preservation of fossils. The Cornubian depositional basin was restless, and received different kinds of sediment from different source areas throughout much of its existence. Few instances of reliable estimates of undeformed thicknesses can be given because of tectonic repetition and folding. Nevertheless, the original thickness of the pre-Permian is estimated at around 5 km: a thickness increased three times by folding and nappe formation.

Much of the framework was understood 25 years ago; subsequent work has focussed on the sedimentology of the Palaeozoic rocks and models proposed for their origins. More significantly advances include the dating and distinctive origins of the Devonian rocks of south Cornwall, the dating and order of the clastics and carbonate facies in north Cornwall and south Devon, the origins of the Culm formations and the later history of the "Culm Basin". Much of the debate about the tectonic implications of this work is summarised by Edmonds *et al.* (1979) and Durrance and Laming (1982). The contemporaneous igneous rocks are discussed below (p.280).

Devonian stratigraphy and sedimentology north of the Culm "synclinorium" has been studied by relatively few workers, amongst whom have been Goldring (1971), Holwill (1962, 1964), Simpson (1962, 1964), Tunbridge (1980, 1984, 1986), Webby (1965a, 1965b, 1966a, 1966b) and the officers of the Geological Survey (Edmonds *et al* 1985, Freshney *et al* 1979).

The vigorous deposition of major alternating coarse and fine-grained formations in a sequence of migratory non-marine to marine Devonian transitional environments has now been well documented. Webby (1966a, 1966b) summarised the Middle-Upper Devonian transgression there, and Tunbridge (1983) has proposed a shoreline facies model. A mid-Devonian "Bristol Channel land mass" was present only as short-lived local uplifts along fault-lines.

Devonian stratigraphy in south Devon reveals regional developments of carbonate facies in the Plymouth area (Chandler and McCall 1985) and the Torbay-Newton Abbot area (Scrutton 1977; Selwood *et al.* 1984). Dineley *(i966)* showed that the oldest known strata, the Dartmouth Beds, include fluvial and other non-marine sediments akin to those of the Lower Devonian of the Anglo-Welsh basin (see also Bridges 1969). Richter (1967), Pound (1983) and others have described the tidal and neritic shallow water muds and sands of the Meadfoot Beds and recently it was suggested that the entire carbonate sequence of the Torquay area sits as a klippe upon the Lower Devonian, having been displaced from a south-eastern source (Coward and McClay 1983). No such suggestion has been made for the Carbonates at Brixham (Mayall 1979).

The resurvey of the Newton Abbot Sheet, Selwood *et al.* (1984) led to the recognition of several deformed contemporaneous successions in rocks of Mid-Devonian to late Carboniferous. This style of stratigraphy with laterally equivalent beds displaced in stacked nappes has now been extended across South Devon into North Cornwall as far as the coastline (Isaac 1985; Selwood, Stewart and Thomas 1985). Tectonic transport as well as sediment provenance is from the south with total displacement many kilometres from the depositional sites. The facies are very largely argillaceous with minor carbonates and rare coarser sediments in the Devonian. Fossils are pelagic and indicate deep waters. At Marble Cliffs the late Devonian carbonates are of turbidite origin and are dated by conodont and cephalopod faunas (House 1963; Gauss and House 1972; Kirchgasser 1985).

In the Liskeard area a sequence from Emsian to Upper Frasnian reveals slates and thin bands of volcanics and thin limestones (some correlative to the Plymouth limestone, Burton and Tanner 1986).

The Devonian succession in south Cornwall displays different characteristics, thought to originate from deposition in a fore-arc trough and lying close upon the northern side of a rising mass (Hendriks 1971). Some authors have included the Lizard Complex in this rise (but see below, p. 281). Macro-fossils are here limited to poorly preserved long ranging plants but discovered palynomorphs promise precise correlation. Pre-Devonian fossiliferous horizons have been mapped as faulted blocks in the Roseland peninsula (Sadler 1973) but these are now regarded as olistoliths (Leveridge *et al.* 1984). They may have been derived from a source area to the south during Devonian time by 'wildflysch" -type sedimentation.

The debate about the detailed stratigraphy of south Cornwall has been summarised by Wilson and Taylor (1976) but dispute continues. Rattey (1980) and Leveridge and Holder (1985) have reinterpreted the transition between Mylor and Gramscatho as an olistostrome at the top of the Mylor Slate. A new interpretation of local structure has been offered on the basis of the stratigraphic evidence there. Barnes and Andrew (1986) relate the stratigraphy and structure to the closure of a small oceanic basin in Devonian time (see below).

At this point it may be appropriate to refer to the opinion that the Start Schists are not Precambrian or metamorphosed earlier Palaeozoics thrust into place from the south (Hendriks 1939). It is held now to be likely that the schists are highly deformed equivalents of the Lower Devonian slates which lie to the north of the steep bounding fault between Hope Cove and Hallsands Bay. Dodson and Rex (1971) obtained mineral ages of 305-255 Ma from the Start Point area and Coward and McClay (1983) propose that local early tectonism has carried the schists over Lower Devonian rocks but were then subjected to further deformation (Hobson 1976) until a back-fold developed in the centre of which the schists

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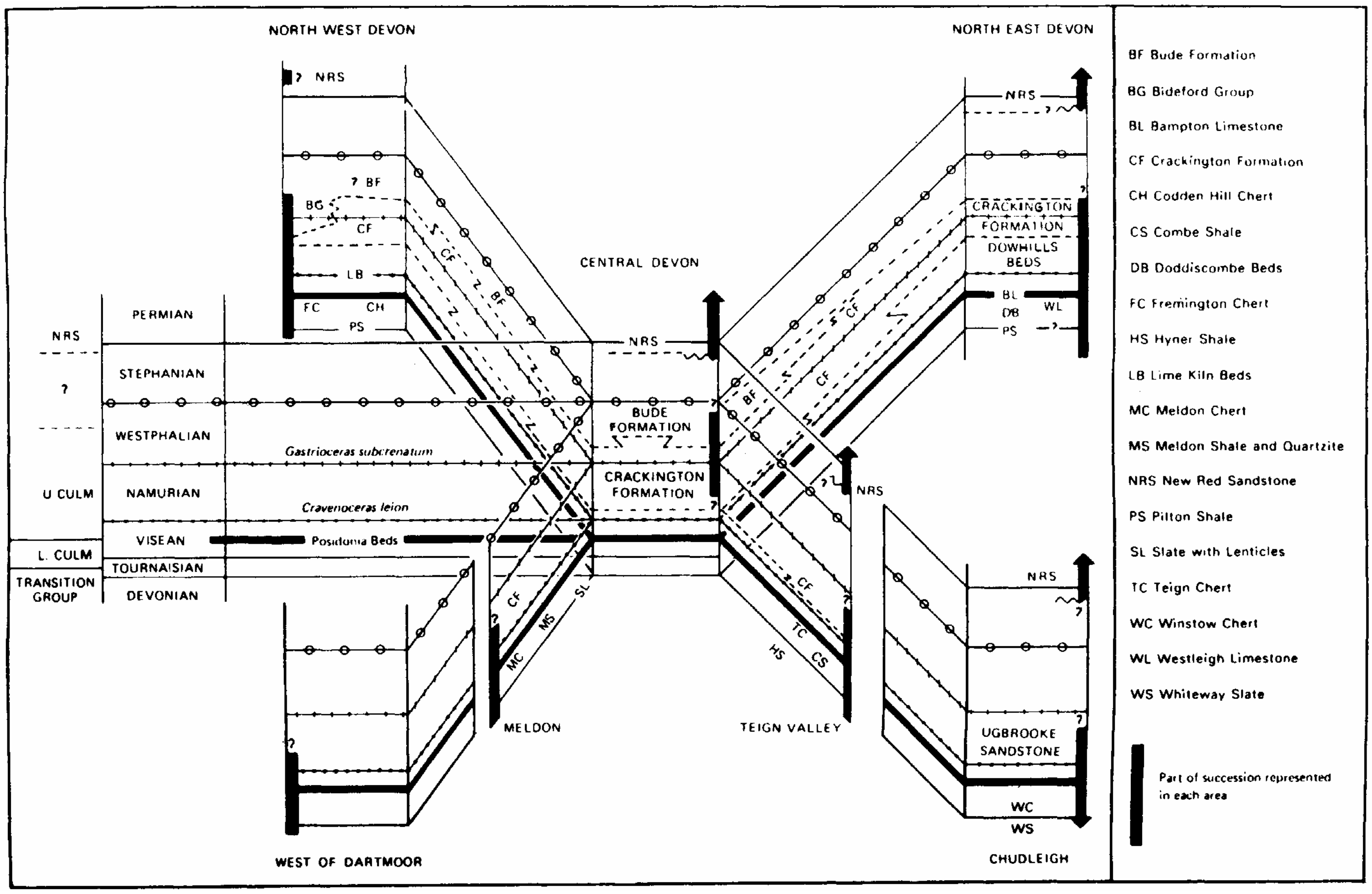

Figure 2. Stratigraphic variation within the Culm Measures in Devon (from Durrance and Laming, *The Geology of Devon,* with permission.

now lie.

The Lower Carboniferous rocks in Cornubia display different facies in the north and the south of Devon and in Cornwall. Dinantian dates have been confirmed for many formations despite the scarcity of fossils (George *et al.* 1976). In North Devon the rocks are deep-water, sandstones, shales, cherts and limestones (Goldring 1962). Prentice and Thomas (1960) and Matthews and Thomas (1974) identified significant ammonoid and conodont faunas near Barnstaple and also at Westleigh where occur turbidite limestones derived from the early Carboniferous shelf to the north.

In south-east Cornwall the Dinantian contains the shallow-water Crocadon Sandstone - a deltaic deposit and flysch sediments deposit farther north. This is said to imply a rising southern source area (Whiteley 1984). Flysch prograded north to reach North Devon by Namurian times. In the south the Namurian flysch is proximal (Ugbrook Sandstone and black shales) whilst to the north distal flysch (Crackington Formation) shows east-west transport axial to the Culm Basin.

Such deposits are best known in the cliffs of the west coast (Freshney *et al.* 1972; Freshney and Taylor 1972) are now interpreted as the clastic infilling of a marine basin by pelagic and turbiditic materials to the point where brackish water and even non-marine conditions prevailed. "Cyclic" sedimentation has been described in several of the formations (see de *Raafet al.* 1965; Walker 1970). Higgs (1984) believes the Bude Formation to have originated in water less than 200 m deep for most of the time, and largely non-marine. Marine bands (with ammonoids) represent short-lived incursions of the sea in a large lacustrine basin. Matthews (1977) regarded the non-marine aspect of the Bude Formation as the most conspicuous feature distinguishing the Cornubian part of the Variscan belt from the German. He thought it had been influenced by the growth of the granite batholith.

The highly variable and in part locally derived Permo-Triassic formations in Devon were previously continued westwards over much of the tectonised Palaeozoic terrain, smothering a highly irregular surface (Laming 1966, 1968). Different depositional environments have been recognised - screes, flash flood fans and sheets, aeolian sand dune and water-channel deposits. Faulting contemporaneous with rigorous erosion and deposition in, for example, the Exeter area may have been local faulting exercising control on lateral variations in the sand accumulation (Bristow and Scrivener 1984). Apart from isotopic dates from the lavas and a few palynomorph determinations dating of the New Red

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Sandstone rocks west of the Exe is only approximate. Extensive sampling for palaeomagnetic work has been carried out (Clegg et al. 1957; Zijderveld 1967).

Summaries of the Cretaceous rocks of south Devon appear in Rawson et al. (1978), and in Durrance and Laming (1982). During the revision of the Newton Abbot sheet new discoveries of Cretaceous rocks were made (Selwood et al. 1984). The Haldon-Newton Abbot outcrops show some variation in the Upper Greensand thickness (Durrance and Hamblin 1969), perhaps as a result of penecontemporaneous Albian-Cenomanian folding (Hart 1971). The composite nature of the coarse (Eocene) gravels that surmount the Greensand has been documented in detail (Selwood et al. 1984). The relationship of the Aller Gravels to the Haldon Gravels is not absolutely clear but both are fluvial and include flints derived from the residual Tower Hill Gravel and pebbles from the aureole rocks in the high ground west of the Bovey Basin (Hamblin, 1974). Hamblin (op. cit.) believes that the Aller Gravel is younger than that of Bullers Hill and probably formed only when the Bovey Basin began to subside.

The remarkable basins with Oligocene infills, of which the Bovey Basin is the principal, have been the subject of both academic study and economic exploitation (Edwards 1976; Freshney et al. 1979; Selwood et al, 1986a, 1986b; see also Nicholas 1980). These deep NW-trending fault-associated basins contain ball clays that were previously held to derive directly from the stripping of kaolinite from local granites. This view has been challenged and a persuasive hypothesis of derivation from local shale weathering has gained favour (Bristow 1968, 1977). Only the top 300 m or so are known in a sequence said to approach 1245 m (Fasham 1971). The ball clay, sands and lignite assemblages at Bovey, Petrockstowe and elsewhere appear to represent lacustrine, swamp and fluvial environments with a local warm to temperate climate flora.

Igneous Rocks

The great variety of igneous rock bodies in the peninsula resolves into the suites of Devonian and Carboniferous volcanics, the granites of the Cornubian batholith and associated minor acidic rocks, and the post-orogenic Permian volcanics. There are also the rocks of the Lizard Complex and the Lundy granite.

Rapid analytical methods, petrological and mineralogical study and isotope techniques have advanced the interpretation of all groups of crystalline rocks. Radio-metric dating has been applied to many outcrops (Dodson and Rex 1971; Fitch et al. 1984; Styles and Rundle 1984). Geophysical data has led to improved models of the batholith shape, extent and possible origins (Bott et al. 1958, Bott and Scott 1964; Brooks et al. 1983).

Variscan magmatism covered 110 Ma (Floyd et al. 1983) and the pre-orogenic volcanics were associated with basinal subsidence of the Cornubian part of the Rhenohercynian belt. These were mobile linear areas of subsidence, moving northwards with time (Matthews 1977; Richter 1965). The abundance of extrusive materials generally increased from early Devonian to early Carboniferous times but came to an abrupt halt in the Visean.

The Devonian and Carboniferous volcanism in our region represents a widespread phase of essentially submarine activity associated with high-level instrusion (Floyd 1982, Floyd et al. 1983). A vast amount of analytical geochemistry by Floyd and co-workers relates these volcanics to the tectonic development of the Cornubian area. The rocks do not resemble modern oceanic basalts in terms of chemical composition abundance or stratigraphic occurrence. Tholeiites prevail, but have many characteristics of continental or intra-plate volcanics.

Interest in the granite outcrops has remained high, several authors commenting on the discordance of these to the strike of the country rocks. A major review by Exley et al. (1983) notes the large size of the underlying batholith and the steeply inclined walls. Six main petrographic types of granite are known: extensive mineralisation and post-magmatic alteration affect all types. Isotope study suggests the source of alteration material of the granite to have originated high in the crust.

Watson et al. (1984), comparing Caledonian and Cornubian granites, noted the close similarity of composition, the steep walls and flattish roof. They observe that the exposed granites seem to occupy spaces "opened by piecemeal or wholesale subsidence of blocks of envelope rocks". (Geophysical data indicate a rather uniform composition to a depth of 10 km.) Many intrusive bodies may be present, emplaced within and alongside one another. Shackleton et al. (1982) have postulated that the batholith is rootless, being fed in from the south and perhaps originally approaching to within I km to the surface (see also Edwards 1984). The ensuing high heat flow in south-west England has been due to the presence of radioactive isotopes of U, Th, and K in these granites.

Geochemistry of the granites, the minor associated intrusives and of the metamorphic aureoles has made many advances (Hawkes and Dangerfield 1978; Exley et al. 1983; Floyd 1982). The age of these rocks is stated as 269 ± 8 Ma and associated with this late stage are mineralisation and post-magmatic alteration. Space does not allow reference to the progress in this field; see, however, Badham 1980; Edmonds et al. 1975; Exley and Stone 1981; Hawkes and Dangerfield 1978).

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Post-magmatic volcanics are confined to two groups - the Plymouth area rhyodacitic lavas and intrusives and the basalts and lamprophyres of the "Exeter Volcanic Suite or Series". The volcanism seems to have been confined to fault-bound troughs within the Palaeozoic during uplift after the orogeny. These rocks have proved of considerable interest geochemically. See Cosgrove 1972; Cosgrove and Eliott 1976; Exley et al 1983; Knill 1969. The Dunchideock (Exeter) volcanic rocks have been dated as 281 ± 11 m.y. (Miller and Mohr 1964) and the lavas to the east of Exeter as 278 ± 6 m.y.; these ages seem to approach the Carboniferous rather than the Permian.

The Lizard Complex, long recognised as an "exotic" assemblage has aroused controversy not yet stilled by an impressive array of technologically well-supported contributions. Opinion is gathering in favour of it being an ophiolitic mass of Devonian age, thrust northwards at an early stage in the tectonic evolution of the Cornubian basin (see the reports from the thematic meeting on the Lizard Complex in the *Journal of the Geological Society of London,* volume 141, 1984; Floyd 1983).

The general view of an obducted ophiolite-bearing nappe contributing debris to the Meneage Formation has been challenged by Barnes (1984). The ages of the gabbro and Kennack Gneiss have been determined as 375 ± 34 Ma and 369 ±12 Ma respectively, which effectively counters the opinion that the Lizard Complex is a slab of uplifted metamorphic basement (Bromley 1973; Matthews 1981) contributing debris during the Devonian period. Badham (1976) offered to this Society the view of a Variscan ophiolite emplaced during the early Devonian. Although undoubtedly ophiolitic in its geochemistry, the gabbro and associated rocks are not strongly analogous to the Troodos or Oman ophiolites. These are now postulated as originating from a spreading centre with dykes provided from a pulsatory magma chamber (Kirkby

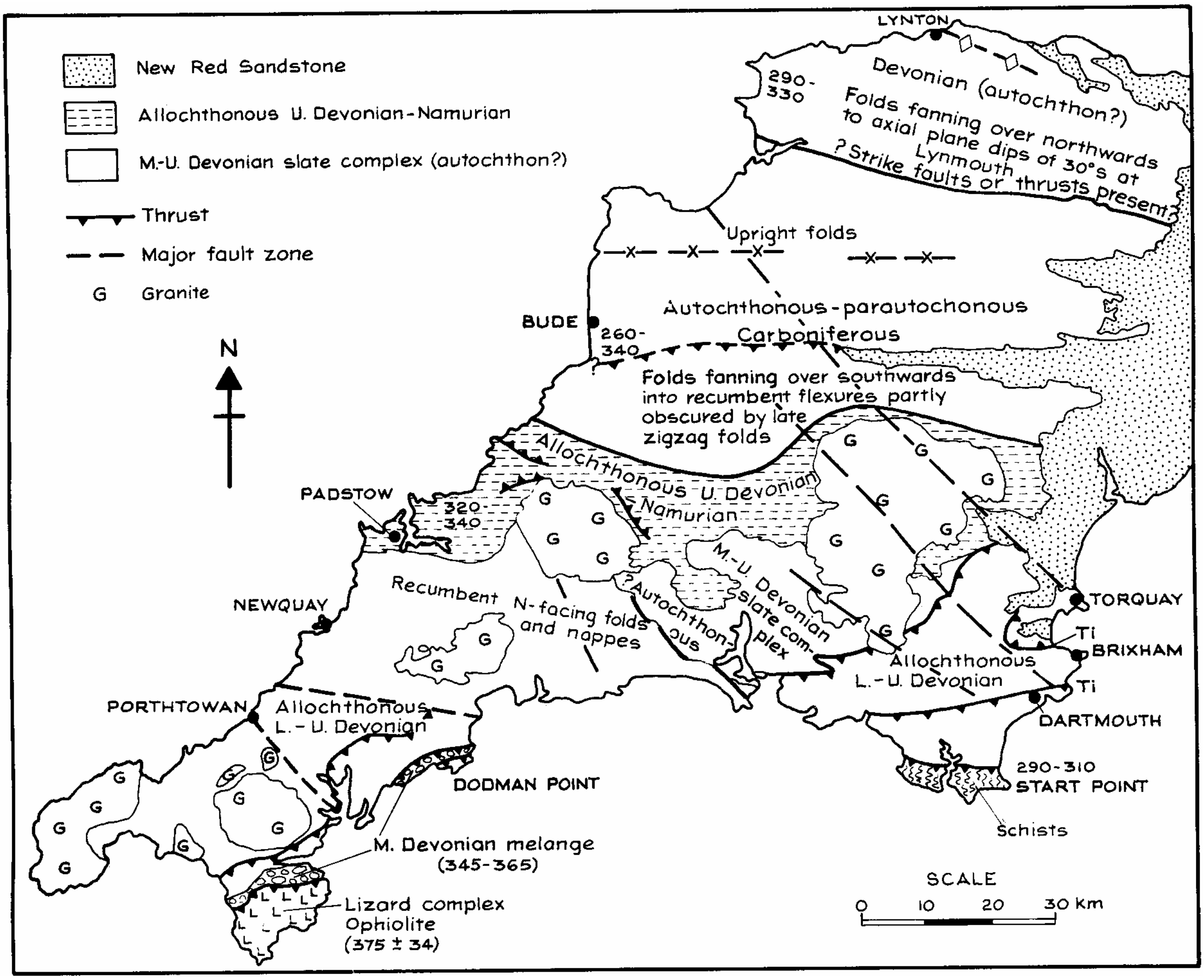

Figure 3. Map of SW England showing major structural features mainly after Isaac (1985) and Leveridge and Holder (1985), with dates after Dodson and Rex (1971). (Ti = thrusts in Torbay area postulated by Coward and McClay (1983)).

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1984). Barnes and Andrews (1986) believe the ophiolite was moved from its original site under conditions of relatively low grade metamorphism and that some 12 km of overburden has been removed. The Mylor-Gramscatho basin may represent parts of the trough which was overthrust by the Lizard nappe from the south.

The Lundy Island granite and its associated minor intrusives (Evans and Thompson 1979) seem out of place in association with Cornubian geology. Edmonds *et al.* (1979) maintain that the fault control along NW-SE lines was extended during the Eocene at which time intrusion of both the Lundy granite and the associated doleritic dyke swarm took place.

Structural Geology

During the last quarter century the suspicions that the structure of Cornubia is more complicated than existing maps suggested has been amply vindicated. Detailed study of cliff sections and the mapping of lithological and tectonic units have led to new local interpretations and regional models. Dearman's early papers (see 1969) indicate the state of structural geology here almost 25 years ago; Hendriks (1939) had contributed an overview of Cornish structures. Significant advances since then have been related in Hutton and Sanderson (eds. 1984). The division of the region into tectonic zones (Sanderson and Dearman 1973), and the age dating of the rocks have been cautiously accepted. Geophysics has given data on rock distribution at depth both on and offshore. The episodic nature of the deformation has been emphasised, and the roles of early thrusting and late normal faulting have drawn attention. The early start of the deformation is noted by workers in the Culm *(e.g.* Whalley and Lloyd 1986) as well as by those concerned with older rocks.

As in other areas of intensive study there is controversy over both data and interpretations. Recently attention has focussed upon structures in the south and upon facing directions as in the 'Polzeath confrontation' (Matthews 1977; Isaac *et al.* 1982; Hobson and Sanderson 1983). The re-mapping of Geological Survey sheets has provided most of the data for this (Selwood *et al.* 1984; Selwood and Thomas 1985, 1986 a, b; Isaac 1985), Interpretations of regional structures in south-west England (e.g. Shackleton *et al.* (1982) amongst others) emphasise their recumbent and thrust nature and gentle tectonic dip. Translation north-northwestwards is achieved in a model of "thin-skinned" tectonics with a basal décollement dipping gently southwards: the granites are thought to be "exotic", injected from a southern source just above the décollement.

Thrust tectonics in north Cornwall have been long in vogue, but the detailed and substantive evidence for them in Devon has only appeared more recently. In Devon, Annis (1933) mapped a thrust at Chudleigh and Vachell (1963) proposed a nappe in the Torbay area. The recent papers by Isaac and others (1982), Isaac (1985) and Turner (1985) now extend our knowledge of thrust nappe Devonian and Carboniferous successions from the east westwards from Dartmoor to the Cornish coast (Selwood *et al.* 1985).

In the South Hams area Richter (1969) and Hobson (1976), have described structures north of the Start Boundary fault which is generally accepted as a steep northerly dipping fracture (Marshall 1985; Saddler 1974). Shackleton and his colleagues (1982) and Coward and McClay (1983), however, see the Start rocks as the core of the backfold which has been reverse faulted along a thrust plane. They maintain that a Start Thrust exists, thought field evidence is obscure and perhaps controversial. Bott and Scott's (1964) gravity data suggest that the Schists drop sharply southwards by near-vertical faulting. New geophysical evidence had led Doody and Brooks (1986) to propose the Start metamorphic rocks are indeed pre-Devonian basement, their present position being due to travel on thrusts generated within the crystalline basement rather than over a thick sequence of Devonian strata.

Coward and McClay's (1983) contribution to the discussion of south-eastern Devon includes two postulated major thrusts. One of these involves the 'Dartmouth antiform' as a major thrust structure with a 13 km or so displacement northwards; the second underlies a major klippe containing the Torquay Limestones. Smythe (1984) discounts the former thrust and there is now much interest in the possible role of the klippe in the Torquay region. Chapman *et al.* (1984) carry the idea of thrust tectonics further by regarding all of south-west Devon as allochthonous.

The continuity of a 'Lizard Thrust' with the Start Boundary *via* the Dodman area is no longer favoured (Hendriks 1971; Matthews 1977; Doody and Brooks 1986). Rattey and Sanderson (1984) note six phases of deformation in the Devonian north of the Lizard Complex, with strong northward displacement. They see early deformations as thickening the sedimentary pile and causing a northerly migration of the actual basin with the batholith intrusion then occurring during the third phase. Leveridge *et al.* (1984) propose a major thrust nappe occupying much of the country north-east of the Lizard, as did earlier authors.

The pride of place for controversy on the north Cornish coast must be given to the so-called 'confrontation' referred to above. Selwood *et al.* (1985, 1986) found themselves unable to accept the projection inland of the interpretation offered by Sanderson and Dearman (1973). The inland terrain has revealed thin nappes involving northward transport and much of the interpretation depends upon a newly revised and palaeontologically supported stratigraphy. The Dinantian Buckator Formation is now seen as a totally fault-bound slice within the Boscastle Formation and is

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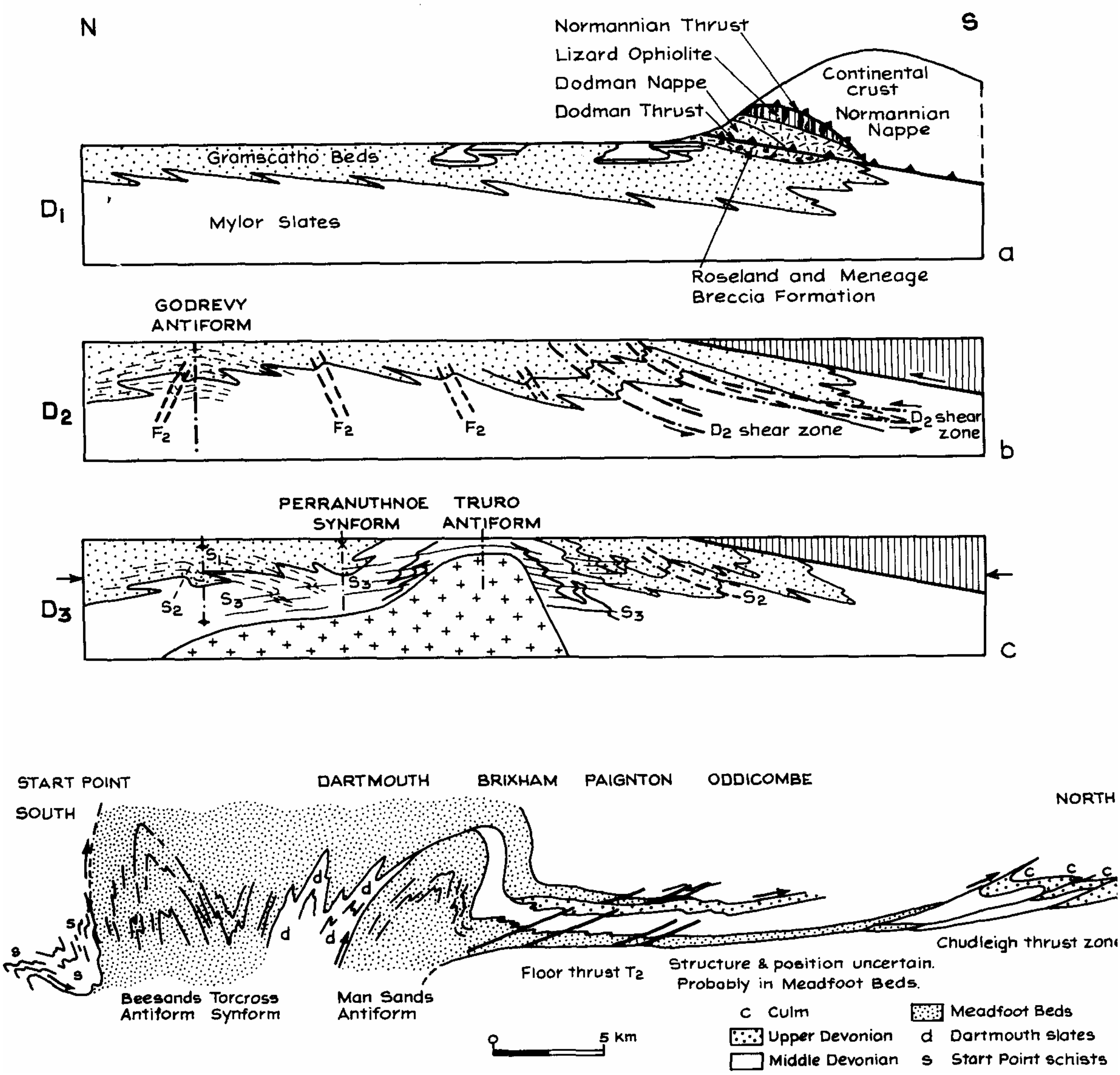

Figure 4. Diagrams to illustrate structure in SW England. D_1, D_2, D_3, The structure evolution of SW Cornwall (after Rattey and Sanderson 1984; Leveridge and Holder 1985). Here the general style of deformation is shown together with the components lying above the Normannian-Dodman Thrusts and including the Lizard Complex (in a). These components are indicated by simple vertical shading in b and c.
The lowest section shows the postulated structure for S. Devon after Coward and McClay (1983).

in the north-western forepart of a major nappe underthrusting the Upper Carboniferous flysch basin of central Cornubia. Local southward-facing structures are thus seen as back-folds in a northerly-moved nappe. The Boscastle Nappe is thus at the top of the tectonic pile derived from the south as recognised inland by Isaac *et al.* (1982). Below the Boscastle Nappe lie the Tredorn (Stewart 1981) and Port Isaac Nappes (Selwood and Thomas 1986 a, b), each with distinctive stratigraphy and structure, and each seems to have been displaced northwards into structural depressions lying ahead of the main deformation front (Selwood and Thomas 1986 a, b).

In summary, new developments in the geology of south and west Devon and Cornwall appear to conform to a pattern supporting the idea of Devonian-Carboniferous sedimentation and volcanic activity in a *schwellen* and *becken* setting with olistostromes, volcanigenic debris flows and dolerites. Intensive synsedimentary deformation took place with perhaps a long extensional

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phase preceding one of compressional stress. This follows the pattern known in Germany (Franke and Engel 1982).

So far no extension of this kind of stratigraphy and structural history has been found in North Devon and West Somerset, Durrance (1985) believes that a major thrust may separate the Bude and Crackington formations along the line of the Crediton valley. Other thrusts may exist along lines parallel to this one, but proof of their existence in north and central Devon is yet to be found. The major structure within the Devonian rocks, the Lynton anticline, is well marked by the outcrop of the Hangman Grits (Edmonds *et al.* 1975, 1985). Widespread thrusts have not been found but strike faulting along shale outcrops probably occurs widely (Holwill *et al.* 1966; Reading 1965). Matthews (1974) discounted an "Exmoor Thrust" but favoured the possible thickening of the Devonian northwards along a line of deep fractures to the south of Exmoor. This would account for the local gravity anomaly (Bott *et al.* 1958). In the Carboniferous dips are high and the folds of small amplitudes. A structural control which locally caused rapid subsidence coinciding with periodic rapid Devonian deposition may have ceased in Carboniferous times with a sudden final deepening after which the basin was (ultimately) rapidly and completely filled. A location in the general area to the south of Exmoor for such a control meets the requirement.

Metamorphism

Structural deformation and plutonic igneous intrusion have each brought about metamorphism of Cornubian country rocks. Slaty cleavage is widespread and varies in dip and intensity from place to place. Thermal metamorphism is distributed as aureoles about the granite bosses. K/Ar dating of slates from south-west Cornwall indicates that deformation began earlier in the south with mica recrystallisation beginning in late Devonian time (Dodson and Rex 1971). From there the dates of metamorphism appear progressively younger northwards. In this is seen a similarity to Variscan metamorphism in north-west Europe. Recent work by Robinson *et al.* (Robinson 1981; Robinson and Read 1981) and by Primmer (I 985a and b) throw light upon the temperatures and pressures responsible for the various grades of regional metamorphism and confirms the diachronism from south to north and the influence of a high geothermal gradient of 40º - 50°C/Km. Fitch *et al.* (1984) showed in the vicinity of the Lizard complex that metamorphism had begun before the ophiolite had been emplaced.

Thermal metamorphic effects in the granite aureoles and post-magmatic processes in the batholith are of both academic and economic interest (see Charoy 1981) with work on fluid inclusions indicating new general models.

Geophysics

Seismic, gravity and magnetic exploration has been used in south-west England to elucidate problems of "shallow geology" and also "deep geology" principally the subsurface location of major discontinuities -thrusts and the shapes of igneous bodies. Brooks *et al.* (1983) have given a general review of work on the Variscides. Palaeomagnetic work has also ranged over rocks as diverse as the Lizard Complex (Hailwood *et al.* 1984) and the New Red Sandstone. Offshore investigations have been numerous, the largest perhaps being the BIRPS and the ECORS programmes (1986: see also Day and Edwards 1983).

Gravity surveys by Bott *et al.* (1958) covered interpretations of the granite masses and the Lizard and other investigations have concerned a noted gravity gradient over Exmoor. The possible presence of a major "Exmoor Thrust" has been disputed (see Brooks *et al.* 1983; Matthews 1977).

Seismic studies have profoundly influenced interpretations of structure to considerable depths and the most recent account (Doody and Brooks 1986) is of a long refraction line between the Lizard and the Start peninsulas. Two layers of regional extent are now proposed within the shallow crust. The upper is of Devonian metasediments, the lower lies several kilometres below the Lizard Complex but rises eastwards to occur at or near the surface in the Start and Eddystone areas. While the Lizard rocks are interpreted as a thin surface sliver about the Devonian metasediments, the lower layer may represent true pre-Devonian crystalline basement, 3 km or so below. Thus the question of the age of the Start rocks is revived, though the Lizard-Dodman-Start thrust hypothesis is not. A Start-type basement seems to be indicated north of the Start boundary fault so the Start block itself may have been transported by thrusts within basement rather than Devonian. It remains to be seen whether geophysical methods will help further to resolve the differences between the Doody and Brooks' model and that of Coward and McClay (1983).

Westwards from Land's End gravity interpretations suggest that the Cornubian batholith extends to about 8° W and that the Haig Fras granite is a separate parallel pluton (Edwards 1984).

Plate tectonics

Stratigraphic comparison of the Palaeozoic rocks of Cornubia with those of mainland Europe, especially of the Rhenohercynian fold bell is inescapable and useful (Matthews 1977; Franke and Engel 1983). As yet, no general plate tectonic model of the evolution of the Variscides has been very widely accepted, though many have been offered. Several basic questions remain to be resolved before such a model can survive. The existence and extent of oceanic crust that may have been implicated, the position, number and activity of

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subduction zones and the nature of the collision (should one have actually taken place) are matters still to be verified.

Undisputed Devonian oceanic crust in Cornubia has not been located· Floyd (1982) concluded that basin transformation began in early Devonian time and ended in Late Devonian before complete continental separation occurred and oceanic crust could appear. Leeder (1982) regarded deposition in "Cornubian' waters as occurring in a back-arc basin, owing its location and origin to "continental crustal 'oceanisation' processes" caused by the Ligurian/Bretonic subduction. While Floyd *(op. cit.)* postulated a shallow northward-dipping subduction zone located south Of the Moldanubian Zone, Leeder suggested that subduction took place in south Brittany, 160-350 km south of the peninsula.

Several authors have favoured a gently dipping subduction zone arid the influence of large strike-slip faults or dextral shear as a Fundamental feature of the Hercynian orogeny. The ubiquitous dextral faulting found throughout south-west England is matched by equally pervasive faulting in European massifs. Badham (1982) regards the Variscides of Western Europe as the result of dextral interaction between the continental European and African plates and the several microplates between them.

The importance of one such major fault is emphasised by Holder and Leveridge (1986b) who note that tectonic continuity between Germany, Belgium and Cornwall was strong until the inception of a dextral transcurrent fault in the Late Carboniferous. It runs from the Bristol Channel towards Portsmouth, Dieppe and south-eastwards to skirt the southern end of the Vosges and Black Forest. Subduction zones previously continuous across this fault occurred (i) along the Mid-German Crystalline Rise to run close to the south Devon and South Cornwall coast and (ii) between the Saxothuringian and Moldanubian Zones continuing near the present north coast of Brittany.

Whatever the configuration of continental crust and the possibility of microplates taking a significant role in the ultimate pattern of the orogenic structures, it is clear that the Cornubian basin was as tectonically active through-out the period Siegenian-Westphalian as any other within the great Variscan trough. A period of extension at least until early Late Devonian time brought about basin subsidence in an area from Exmoor to South Cornwall. Vigorous continental deposition persisted intermittently in North Devon but to the south all was marine with synsedimentary deformation and volcanicity. The southern margin of the basin, rising throughout this time, shed debris northwards and itself moved in the same direction· On the northern boundary the flank of the Old Red Sandstone continent - St. George's Land - remained relatively immobile and fixed, shedding from time to time turbiditic sediment southwards from the shelf. Barnes and Andrews (1986) have used the stratigraphy and structure of the Gramscatho flysch and the Lizard ophiolite to integrate local evolution into that of the Rhenohercynian zone as a whole. Here an intra-continental dextral transform system developed as the local oceanic basin was closed during Devonian time. Beginning as a pull-apart basin, the Gramscatho basin closed at the end of the Devonian and nappes were driven northwards. Foreland basins receiving Carboniferous infill, were sequentially developed and deformed by thin skinned tectonics throughout that period. Holder and Leveridge's (1986a, b) view of the tectonic evolution of South Cornwall relates to the closure of the basin there by southward subduction throughout the Devonian with a subsequent continental collision between the Normanian High and part of Laurentia (Old Red Sandstone Continent, Laurussia). Although deformation continued late into the Carboniferous the collision and closure of the basin was complete by the end of the Devonian rather than later as Ziegler (1982) suggested.

Elimination of the Cornubian basins seems to have been virtually complete by the end of the Westphalian time. Granite intrusion was enacted by then, its paths being perhaps guided by a zone of structural weakness rising northwards (the zone of décollement of Shackleton *et al.* 1982) from the proximity of a subduction zone. but relatively high in the crust.

In common with the remainder of the British Isles, a phase of normal faulting seems to have followed the intrusion of the Cornubian batholith and the climax of orogeny. By this time the Wessex and other basins around the Cornubian massif began to subside, initially as narrow fault-depressions but with vigorous erosion of the uplifted orogenic terrain already having taken place. Permian stratigraphy indicates that this erosion had exposed rocks relatively low in the structural column prior to the formation of the earliest New Red Sandstone. From then on sedimentation spread progressively to cover the Cornubian massif by late Cretaceous time. The palaeogeography of Cornubia from then onwards poses many questions for which there is no space here.

Conclusion

The last quarter of a century has seen a great improvement of knowledge of the outcrop geology of the Cornubian massif and the rocks at its margin· There has been a smaller but important improvement to the sum of deep subsurface geology· While field techniques have not changed greatly, laboratory techniques and methodology have become more sophisticated, more rapid and accurate in their results and data handling has been accelerated by computing aids. Our view of the stratigraphy and structure of the south-west of England has inevitably become cognisant of regional variations and complexities· On the sub-Devonian crust and the influence of structural features within it we have not

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significantly advanced our knowledge but much has been gained by detailed local analysis and consequent comparison with other parts of the Variscan fold belt. German sections and interpretations have been important in this respect but despite all the similarities noted, there is no convincing evidence that there was unbroken continuity in each depositional environment from Germany to south-west England.

There remain areas of controversy, especially in the interpretation of local structural details and their importance to the model of regional deformation and of the region's significance in a major plate tectonic scenario. The next quarter-century may bring about a greater uniformity of view. It is certain that members of the Ussher Society will be as active in contributing to that view as they have been in making the present brief overview possible.

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The Scott Simpson Lecture

Read at the Annual Conference of the Ussher Society, January 1986

The deep geological structure and evolution of southern Britain

A. WHITTAKER

Whittaker, A. 1986. The deep geological structure and evolution of southern Britain. *Proceedings of the Ussher Society, 6,* 291-298.

Maps of the surface geology of Britain yield information that is of great use to deep geology. The presence of unconformities, however, necessitates the use of seismic reflection and deep drilling data to elucidate the 3-dimensional structure of the continental crust in any detail. Recently acquired deep seismic reflection profiles suggest that the British crust is divisible, vertically, into three seismic zones which correspond to tectonic units in southern Britain. Of particular significance is the recognition of thrust faults and planes of decollement in concealed fold belts; detailed analysis of the post-Carboniferous basins shows that these major fractures have re-activated with profound effect upon basin development and subsequent structural inversion.

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1. Introduction

The 1:625000 geological map of Great Britain summarises 150 years of work by the British Geological Survey. It shows the surface distribution and disposition of the major British lithostratigraphical divisions. Study of the map reveals important and significant changes of strike from place to place, one of the most obvious of which is that of the Mesozoic formations when traced from the Midlands area into extreme southern England Across the zone of the projected ESE-trending Variscan Front the strike of these rocks changes from broadly NE to ESE or E-W. Additionally, the outcrop patterns themselves give many clues about the structural disposition and geological history of major stratigraphical units. For example, the western and central parts of Mesozoic-covered Britain show elongate, relatively narrow 'tongues' of Triassic and Lower Jurassic (Liassic) sediments of varying trend commonly bounded by what used to be known as Jurassic axes of uplift.

The map also strikingly illustrates unconformities of varying degree. Particularly obvious are those at the mid-Cretaceous level (late Kimmerian), the base Permo-Triassic level (Variscan), the base Upper Palaeozoic level (Caledonian) and the base Lower Palaeozoic level (Precambrian). The presence of unconformities, separating rocks units of different structural disposition, makes it difficult, under most circumstances, to predict or prognose what is present in the subsurface, down the dip, away from the outcropping unconformity. Without subsurface geological and geophysical information, for example, it would be difficult, if not impossible, to deduce the presence of a major structural entity like the London Platform from surface geological mapping alone.

Using the existing data base it is possible to calculate roughly the amount of surface area of mainland Britain occupied by the major tectonostratigraphic levels of the Mesozoic, Upper Palaeozoic, Lower Palaeozoic and Precambrian, each separated from the other by important unconformity (Fig. 1). Bearing in mind the importance of the Upper Palaeozoic level alone in terms of mineral resources (oil, coal, metalliferous minerals) it will be seen that as detailed a knowledge of the hidden 39 per cent or so as possible is imperative from the economic as well as from the scientific point of view. An example from within the Mesozoic level of southern England will suffice to show the importance of detailed knowledge of the deep geology. Early attempts at oil exploration in southern England concentrated on anticlinal structures deduced from surface mapping and were largely unsuccessful; with the advent of new techniques to examine the sub-late Kimmerian geological structures developed in Jurassic, Triassic and Permian rocks a new era of successful exploration was initiated and the attendant data gathering and interpretation still continues.

2. Deep geological techniques

Prior to the mid-1970s much of the knowledge of the deep, subsurface geology of southern Britain relied upon cores and samples from shallow or intermediate depth boreholes, the potential field geophysical methods and data, and what could be deduced from geological relationships observed and measured at outcrop. Of

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particular importance to deep geology are the results from seismic reflection profiling and drilling because data from these techniques can provide a detailed geological picture of the subsurface.

The Deep Geology Research Group (DGRG) of the British Geological Survey (BGS) has been acquiring seismic reflection data over the last eight years or so by several different routes. BGS holds, confidentially on behalf of the Department of Energy, copies of the oil company seismic profiles but these are commonly only recorded to 3 or 4 seconds of two-way travel time (TWIT). However, there are numerous examples of DGRG collaborating with oil companies to record data to 12 seconds TWTT which gives information to the base of the crust. In addition DGRG commissions seismic parties specially to record profiles to 12 seconds TWTT in areas with outstanding geological problems or with geothermal potential, and also uses in-house BGS equipment for similar purposes occasionally. The results of these surveys, like those recorded offshore by the British Institutions Reflection Profiling Syndicate (BIRPS), are giving a new insight into the deep geological structure of Britain.

Deep drilling in Britain has also changed substantially over the past ten years or so. During the first half of this century relatively deep drilling was undertaken by coal and oil exploration companies and restricted, of course, to prospective areas. Between the 1950s and 1970s, however, BGS carried out drilling for scientific purposes in various parts of Britain usually to solve local geological or geophysical problems and, in places, up to depths of 1100 m or so. In most cases these boreholes were fully cored and, often, comprehensively geophysically logged. They have added much to knowledge of British stratigraphy and geological history by presenting detailed information that could not be gained in any other way. Since the /nid-1970s increasing costs have seen a diminution in the amount of coring but an increase in the quantity and quality of downhole geophysical logging being carried out. In addition to their uses in petrophysics (porosity determination etc.), which are fundamental to many parts of the oil industry, geophysical logs can be used in lithological determination, sedimentological work, depth of burial studies and, of course, in stratigraphical correlation (Whittaker *et al.* 1985). It is clear from the data already available that detailed intra- and inter-basin correlation can be achieved and that in some parts of the geological column (eg. in monotonous red bed sequences) it is easier to identify stratigraphical units from the geophysical data than it is to recognise them by eye in cores. In open-hole, tricone bit drilled sequences, of course, where cuttings samples are of uncertain depth, the only way of delineating stratigraphical boundaries precisely is by using the log data.

3. The structure of the British continental crust.

Since 1979 the DGRG has been acquiring and interpreting deep seismic reflection data and has been responsible for geoscientific operations at several deep bore-holes funded by BGS (NERC), the UK Department of Energy and the Geothermal Programme. Early results from the seismic profiling (Whittaker and Chadwick 1983, 1984 and Whittaker *et al.* 1986) suggest that the continental crust beneath southern Britain exhibits a certain consistency of deep seismic structure and that the profiles, usually, are divisible from top to bottom into three zones based upon reflection character represented diagrammatically in Figure 2.

Seismic zone I is characterised by high-amplitude, well-defined, sub-horizontal reflection events of considerable lateral continuity and extending from the land surface to a level of 1 and 2s TWTT, though in places deeper. Seismic zone 2 extends from the base of zone I to variable levels, commonly 5s TWTT. It comprises ill-defined, incoherent seismic reflections and diffractions of low-amplitude, which, on unmigrated sections appear to be cross-cut by dipping, moderate to high-amplitude, roughly linear reflection events. Zone 3 extends from the base of zone 2 (or, more rarely from the base of zone 1) to a level of about 10 or 11s TWTT. The absence of records greater than 12s TWTT introduces some doubt into the interpretation of the position of the Moho, the base of zone 3. The zone is characterised by moderate to high amplitude, well-defined, sub-horizontal reflection events of marked lateral discontinuity. A similar tripartite division of the continental crust of western Europe is discernible on offshore data acquired by the BIRPS group and on data acquired by the ECORS and DEKORP groups in France and Germany, respectively. The onshore British data are interpreted geologically by

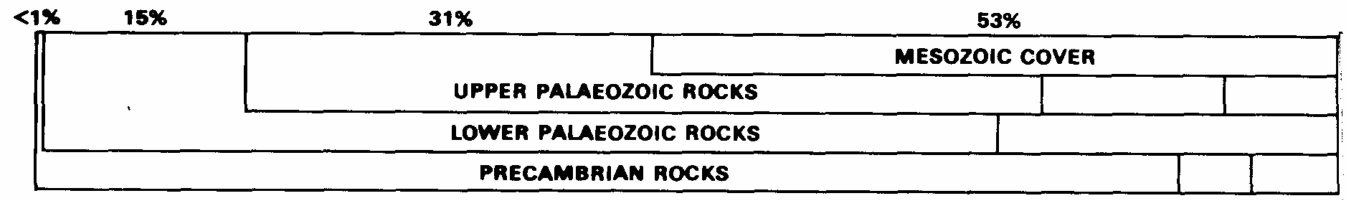

Figure 1. Area of surface exposure of British rocks south of the Highland Boundary Fault. Total land area is 172880 sq.km (100%); Mesozoic cover is 91100 sq.km exposed (53%); Upper Palaeozoic rocks are exposed over 53920 sq.km (31%); Lower Palaeozoic rocks are exposed over 27000 sq.km, (15%); Precambrian rocks are exposed over 860 sq.km (< 1%).

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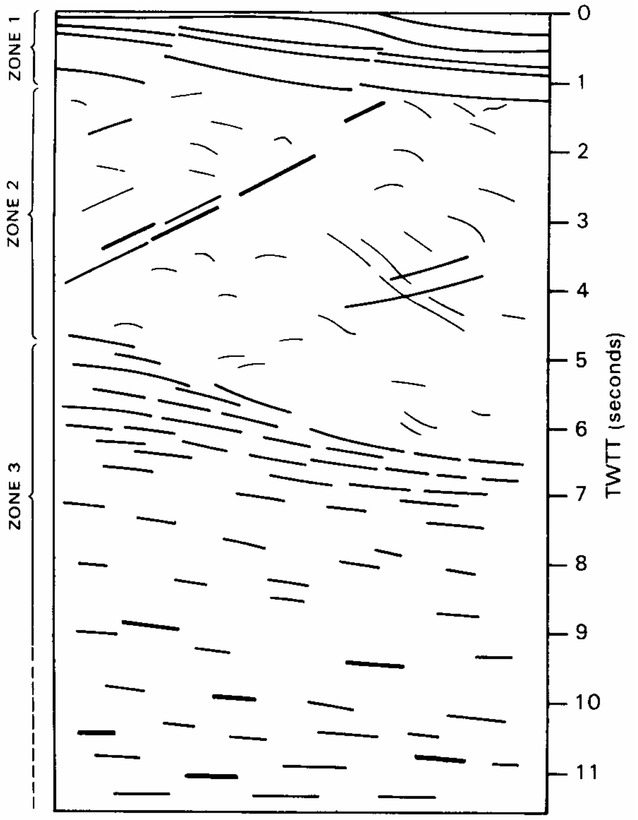

Figure 2. Line drawing of events seen on a typical deep seismic reflection profile.

the DGRG in terms of structural or tectonic units and may encompass rocks of widely different ages. Zone 1 is readily interpretable from abundant seismic and bore-hole data and corresponds with cover rocks of various ages which have not undergone intense structural deformation. Normal faulting is characteristic of the zone, though gentle open flexures and occasional compressional features are not uncommon. In southern England, south of the Variscan Front, zone 1 comprises Permian to Tertiary strata, over the Midlands Micro-craton it corresponds to strata of late Proterozoic to Tertiary age, while in north east England (north of the Midland Microcraton) it comprises cover sequences of Upper Palaeozoic to Cretaceous age inclusive. Zone 2 is associated with reflection events interpreted as major thrust planes or d6coliement surfaces and the seismic reflection character is compatible with the presence of major and minor folds (though limitations of seismic resolution preclude elucidation of detailed fold geometry). Thus the zone comprises, at least in part, complexly folded and faulted strata; south of the Variscan Front zone 2 is the Variscan fold belt, while north east of the Midlands Microcraton it corresponds to the concealed Caledonide fold belt. Zone 3 is present everywhere, apparently, and forms the lower crust. Its lower boundary, the Moho, does not give rise to a well marked, continuous reflection event but is usually indicated by more prominent., discontinuous events between 9.5 and ll.5s TWTT on most deep reflection profiles. The upper boundary of zone 3 is much more clearly marked and must be of considerable geological importance because for the most part structures within the overlying zone do not pass down into it; it varies in level from about 6s TWTT on the south coast to a high point in the vicinity of the Variscan Front, but falls to the 6s TWTT level again in north-east England. Geological interpretation of the reflection events of zone 3 is speculative; only in the vicinity of the Variscan Front may they be Present at drillable depths of 10 or 11 kms or so. Seismic zone 3 has been tentatively identified as metamorphic or crystalline basement characterised by ductile deformation fabrics. Modelling studies (Fuchs 1969) suggest that similar events can be produced by interference of reflections from thin (approx. $\lambda/4$; say c. 250 m-thick) alternating high and low velocity layers of variable thickness, Such features may be due to cumulate layering (Eisner and Lushes 1983) or, as favoured here, granulate facies gneisses showing a dominantly horizontal tectonic fabric. A preliminary sketch cross section through the crust of southern Britain is shown in Figure 3; it incorporates seismic refraction, reflection and borehole data from numerous sources and shows the P-wave velocity structure of the crust

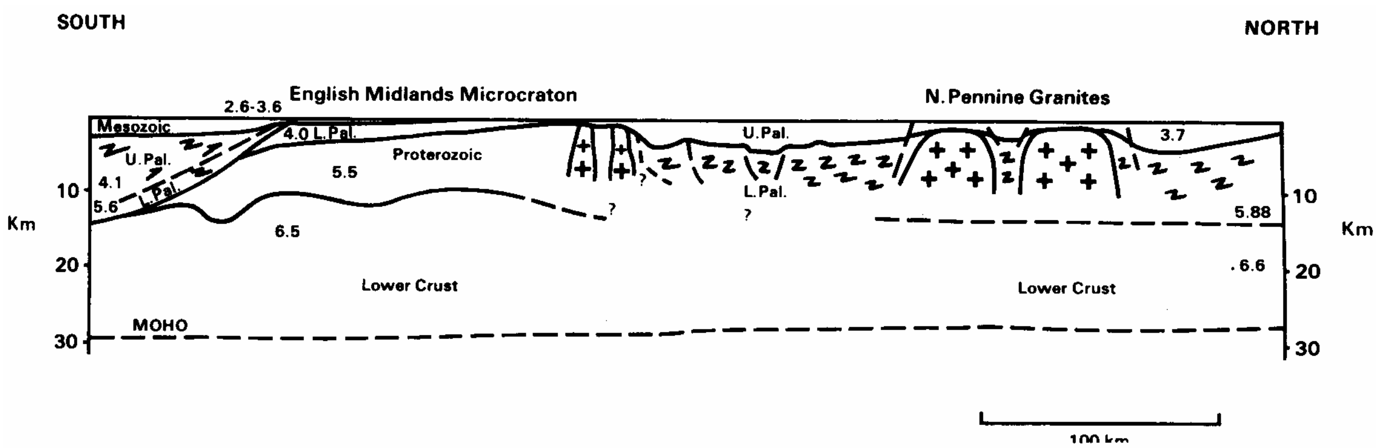

Figure 3. Sketch section through the crust of southern Britain based on data from many sources and showing P-wave velocities in km/s^{-1}

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4. Examples of the concealed sub-Mesozoic geological structure

The study of lower crustal structure by the seismic reflection method ideally requires horizontally long profiles over tens of kilometres of line length. This is much less expensively accomplished offshore than onshore and explains why there are so few long profiles onshore in Britain. However, the seismic reflection method is unrivalled in elucidating mid- and upper crustal tectonic problems in areas where there is concealment of older rocks by younger; indeed, drilling and reflection profiling are the only techniques capable of tackling what may have been outstanding geological problems for decades, in areas where the surface geology is relatively well understood.

One example from southern Britain concerns the western margin of the Worcester (Severn) Basin, where the basin infill rocks at outcrop (mainly Mercia Mudstone Group) are in contact with much older (Lower Palaeozoic or Precambrian) rocks in the vicinity of the Malvern lineament. Uncertainties about the depth of the basin, the nature of its fill, the presence of pre-Permian sediments, the depth to basement and the nature of the contact at the western margin were much debated in the geological literature for many years. Recently acquired seismic data (Chadwick 1985) have clarified the structural relationships to demonstrate the presence of a major fault forming the western boundary to the Worcester Basin. Seismic lines BGS 84/01 and 84/02 illustrate the main features well but have not yet been described in detail (Fig. 4). The geology of the eastern part of the section is constrained by the released Dempsey No. 1 Well (SO 861 493), located just south of Worcester, which proved a sequence similar to that drilled at the nearby released Nether ton well. Dempsey No. 1, drilled in 1979 by DGRG on behalf of Department of Energy's Petroleum Engineering Division, proved 348 m of Mercia Mudstone Group overlying 1018 m of predominantly arenaceous sediments of the Sherwood Sandstone Group, overlying in turn some 938 m of Permian Mottled Sandstone. An important unconformity penetrated at a depth of 2305 m separated the overlying Permo-Triassic sequence from

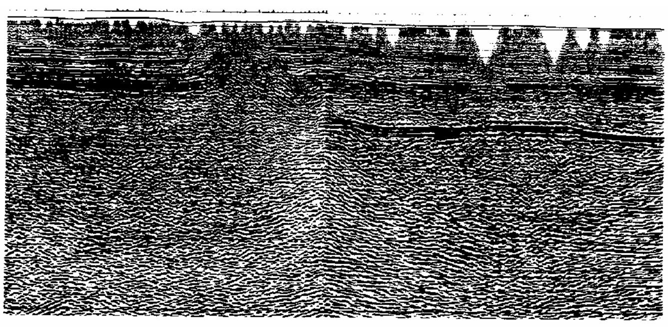

Figure 4. BGS seismic lines 84-01 and 84-02 showing a roughly E-W section across the Malvern lineament north of the latitude of Worcester (for discussion see text). The record length is approximately 4sTWTT.

purplish red and green tuffs and agglomerates assigned to the Precambrian and penetrated to the final depth of 3011 m. The geophysical log data show that the prominent, high amplitude seismic reflection event at approximately 1.6s TWTT at the eastern end of the section represents the upper surface of the ?Precambrian which can be mapped beneath large parts of the Worcester Basin area. The Lower Palaeozoic shelf sequence of the western part of the seismic section, west of the Malvern lineament is constrained by the nearby, released Killington No. 1 borehole which proved about 564 m of Downtonian sediments resting upon almost 1158 m of Silurian strata including the Aymestry Limestone, the Wenlock Limestone and the Woolhope Limestone. The Precambrian basement, located at and below the level of approximately 0.8s TWTT at the western end of the seismic profile shows faint, apparently easterly-dipping reflection events Which may represent thrusts.

The longest, onshore, deep seismic reflection profile in Britain is located between Wiltshire and Dorset. It crosses the Variscan Front, separating the Variscan foreland to the north from the Variscan fold belt to the south and was described in detail by Kenolty *et al.* (1981) and Chadwick *et al.* (1983). The shallow Mesozoic part of the sequence is understood well and displays major growth faulting and the accompanying thickness changes (Fig. 5). The sub-Permian, Phanerozoic sequence is constrained by several deep boreholes which have penetrated beneath the Mesozoic cover. Particularly noteworthy is the released Cooles Farm [SU 016 921] well which proved Permian and Mesozoic strata to a depth of 1205 m resting upon a predominantly argillaceous Lower Palaeozoic sequence with basal Cambrian sandstones proved to the terminal depth of 3513 m. Data from the well enable calibration of the northern (Variscan foreland) part of the section and prove that the base of the Phanerozoic sequence hereabouts coincides with the prominent, high amplitude reflection event at about 2.3s TWTT at the northern end of seismic line IGS 79/01. Everything below this reflection event at the northern part of the section is therefore part of the Precambrian basement (see Fig. 5). South of the vicinity of Calne the geoseismic section shows the Palaeozoic rocks to be over-folded, with steep or overturned northern limbs and long, relatively gently dipping southern limbs. This interpretation is based upon regional geological considerations and upon borehole dipmeter data, plus reflection events near the extreme southern part of the seismic section. The important sub-Mesozoic anticlinal culmination between Devizes and Maiden Bradley is proved by the released Shrewton No. 1 Well [SU 031 420] which proved a 1743 m-thick Mesozoic sequence overlying an unbottomed 1251 m-thick Tremadoc sequence to the terminal depth of 2994 m. The Tremadoc proved in this borehole comprises a rather uniform succession of medium to dark grey laminated siltstones, with sub-ordinate mudstones and pale greenish grey very fine-grained sandstones all, for the most part, relatively

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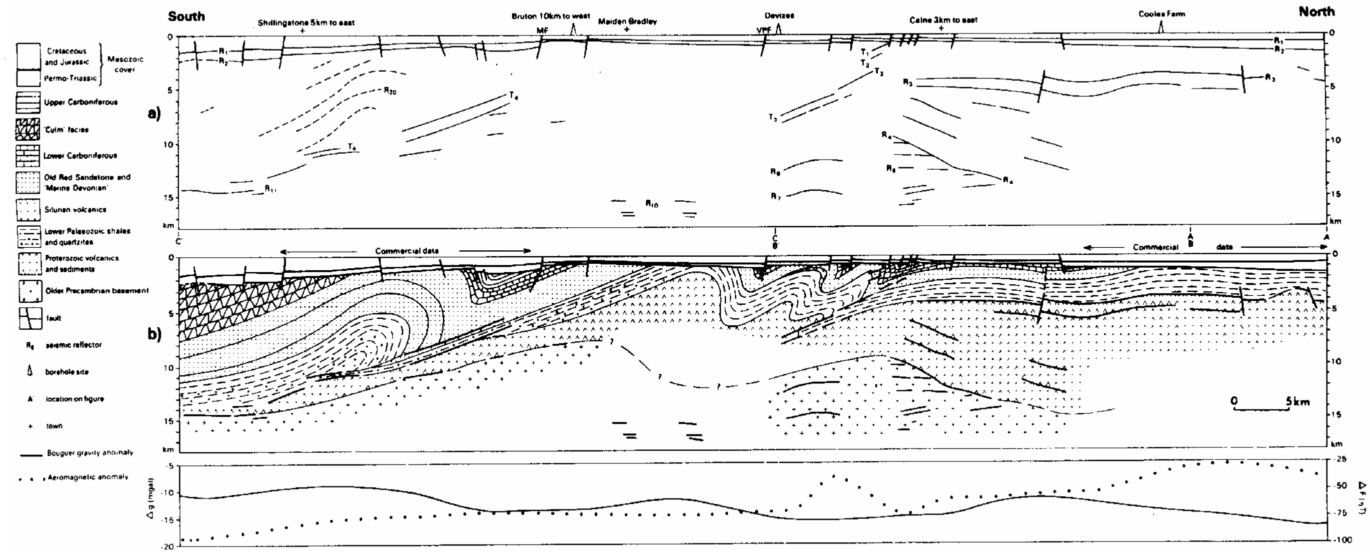

Figure 5. Roughly N-S geoseismic section across southern England.

undisturbed except for fracturing and faulting. The beds contain acritarchs of undoubted Tremadoc age and show a consistent 20º to 40º dip in a south or SW direction. The identification of important, southerly-dipping, high amplitude reflection events in the southern part of the section as thrust faults, is discussed by Chadwick *et al.* (1983). On this section, two major thrust zones are recognised dipping between 22º and 27º with azimuth averaging 160º. The interpreted thrusts are mappable in the subsurface over considerable areas and similar reflection events are seen on other British onshore and offshore data. Although these British thrust-like reflectors have never been proved by drilling, similar events on seismic reflection data from the Variscan Front zone of northern France and Belgium (including the 'Midi' overthrust) have been proved by deep drilling. In the onshore British case, the important thrust-like reflection events underlie major growth faults in the cover sequence. The interpretation of the deeper parts of the Wiltshire/Dorset geoseismic section is necessarily speculative because there is no 'hard' geological data to use as control. However, study of the character of the reflection events and their relationships to each other led Chadwick *et al.* (1983) to distinguish interpreted sub-Cambrian reflection events in terms of a 6 to 11 km-thick sequence of late-Proterozoic rocks in the northern part of the section, thinning southwards. This suite of reflectors is not in seismic conformity with the deepest, sub-horizontal reflectors which were correlated with older crystalline basement. How this basement may relate precisely to the Pentevrian basement of north west France to the south, or to the Lewisian basement of north west Britain to the north is not clear from the available data; it may be that the basement 'core' of the Midlands Microcraton was once part of the Fennoscandian - East European Platform, but this is speculative. Judging from the interpreted Wiltshire/Dorset geoseismic section it seems that the topmost surface of this mid-lower crustal Precambrian basement has a considerable subsurface topography of several kilometres amplitude; in cross section the waveform of the surface is asymmetrical with gentle southward slopes and more steeply inclined northward slopes. This might be connected with the development of southerly inclined thrust ramps in the higher part of the section. It is interesting to note that the Midlands Microcraton or Variscan Foreland of the section is underlain by a considerable thickness of Proterozoic rocks and that this tectonic unit has been relatively stable since Proterozoic times and relatively buoyant since late Palaeozoic times.

5. Post-Carboniferous structure and evolution of southern Britain

The post-Carboniferous geological structure and evolution of southern Britain is known in considerable detail from surface geology, drilling and seismic reflection profiling. As part of its contribution to the BGS 150th anniversary celebrations the DGRG compiled and published an atlas of onshore sedimentary basins in England and Wales, describing the post-Carboniferous tectonics and stratigraphy (Whittaker [Ed.] 1985). In addition, BGS, on behalf of the Department of Energy, extended some of the work to include the offshore areas of southern Britain (Smith, 1985 b + c) with a geological map and contours of the pre-Permian surface updating and extending the work of Wills (1973) and Dunning (1966) respectively.

The concealed Variscan fold belt beneath the Mesozoic (including Permian) cover of southern England consists, at subcrop, mainly of rocks of Devonian to Carboniferous (Dinantian) age (Smith 1985a) with E-W structural grain. Near the margin of the Variscides (southern edge of Midland Microcraton), however, are subcropping areas of Cambro-Silurian rocks. It is unusual for more than one Palaeozoic geological system

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to be penetrated in any single borehole and seismic reflectors indentifiable as stratigraphic interfaces, in general, cannot be confidently identified. However, long, gently south-dipping high amplitude events are present in the concealed Variscan fold belt and these are interpreted as thrust faults. The concealed Midlands Microcraton, at the level of the pre-Permian surface comprises Precambrian to Carboniferous rocks with a structural grain which trends N-S in the Worcester Basin area but swings gradually to a NW trend eastwards over the London Platform proper. Lying above the Precambrian of the Worcester Basin is a subcropping sequence of Palaeozoic rocks which youngs towards the east and culminates in the concealed Westphalian beds of the Oxfordshire and Berkshire areas. North of the London Platform the Mesozoic cover sequence in NE England conceals a gently folded and faulted cover sequence of Upper Palaeozoic rocks with NW or NNW-trending structural grain. The combined Mesozoic and Upper Palaeozoic cover sequences hereabouts are floored by strongly folded Caledonian 'basement'.

The present-day structure contours of the pre-Permian surface in broad outline show a distribution of basins and highs similar to those of Kent (1949) and Dunning (1966). However, in detail the more recent maps depict several sub-basins within the major downwarps of western and southern onshore Britain. In particular the Wessex Basin of southern England contains several important depressions of the present-day pre-Permian surface beneath mid-Dorset (the Dorset Basin), beneath the Vale of Pewsey area (Vale of Pewsey Basin) and beneath the western Weald (the Weald Basin). In terms of the Permian and Mesozoic sediments a distinction can be drawn between the relatively undisturbed and shallowly depressed eastern England Shelf (the western margin of the southern North Sea Basin) and the much more deeply depressed and structurally complicated western and southern basins.

The atlas of onshore sedimentary basins presents a series of depth contour and isopachyte maps of the major Permian, Mesozoic and Cenozoic lithostratigraphical units with accompanying textual description of the stratigraphic history, sedimentation and tectonic setting. From the maps it is possible to see that from Permian to late Jurassic times the area of sedimentation increased from the west, north-west and north encroaching on the London Platform by overlap and overstep. The southern and western basins are associated with major growth faulting so that grabens and halfgrabens, with associated synsedimentary thickening of beds, are characteristic features of the earlier basin development. Zones of active normal faulting migrated eastwards across the southern region and the Wessex Basin developed as a series of fault-bounded, asymmetric half grabens (see Chadwick 1985, Holloway 1985, Smith 1985a). In late Jurassic and early Cretaceous times the area of deposition contracted or shrank and two distinct depositional provinces were established separated by an emergent London Platform. The Wealden Beds of southern England are in most places conformable with the underlying Purbeck strata but in places overlap Jurassic formations. Great thicknesses of strata were laid down in the Weald Basin (and in the Central Channel Basin), but active faulting was restricted to Ryazanian and Valanginian times. Sedimentation in late Jurassic and early Cretaceous times was greatly influenced by the interaction of global sea-level and local tectonic subsidence rates and in Hauterivian and Barremian times most of southern Britain was subjected to erosion and uplift resulting in the late-Kimmerian unconformity. Mid-Cretaceous times (Aptian - Albian) saw the Lower Greensand transgressing across the late Kimmerian erosion surface overstepping westwards and northwards onto progressively older formations; a second transgressive phase accompanied the deposition of the Gault and Upper Greensand. Upper Cretaceous subsidence in southern Britain 'was of a regional nature with little significant contemporaneous normal faulting; the end-Cretaceous erosional episode (probably caused by a major fall in sea-level) removed large amounts of Chalk. At the end of Cretaceous times the Permian and Mesozoic basins of onshore Britain had attained their fullest development, with maximum sediment fill estimated to range from 4.5 km in the Cheshire Basin to 3.5 km in the Worcester Basin and 3 km in the Weald Basin. The extensional tectonic regime characteristic of Mesozoic basin development hereabouts was about to give way to compressional tectonics and associated basin inversion during Tertiary times.

Although subsidence continued into mid-Tertiary times, sedimentation was restricted by a decreased area of deposition due to the end-Cretaceous fall in global sea-level. Clastic sedimentation on any scale was generally restricted to the southern and eastern parts of England with a lack of significant normal faulting. However early Tertiary sediments in the west are present but restricted to deep fault-controlled basins such as those at Bovey Tracey and Petrockstow. The Alpine orogenic movements, culminating in Miocene times, and associated with continental collisions to the south and southeast, produced a roughly N-S compressive stress field in southern Britain which gave rise to inversion of major parts of the area. Those areas underlain by Mesozoic basins became anticlines, or highs, while those areas which were Mesozoic highs became synclines. Thus the Mesozoic Weald Basin was inverted to become the Tertiary Weald Anticline and the Mesozoic Central Channel Basin inverted to become the Tertiary Central Channel High. Conversely, the Mesozoic Cranborne-Fordingbridge High inverted to form the Tertiary Hampshire Basin and the Mesozoic London Platform became the Tertiary London Basin. The development of these inversion structures was controlled by easterly or ESE trending belts of reverse faults and folds (monoclinal or anticlinal flexures)

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which bound the major basins and highs. The reverse faults and folds were formed by compressive reactivation of underlying Permian and Mesozoic normal faults. In some cases precise reactivation of an individual Mesozoic extensional normal fault can be demonstrated (eg. the Mere Fault); in such a case the sense of displacement on the fault changes up the sediment pile from normal to reverse. In other cases the reverse fracture follows a new course beneath the major Mesozoic normal extensional fault.

It was noted earlier that the major, normal Mesozoic growth faults overlie thrusts in the more intensely deformed older rocks of lower Structural levels. It is thought that these major syn-depositional normal faults are controlled, in many places, by reactivation of the thrusts within the brittle crust of seismic zone 2. From this it clear that there have been several reactivations of the same lines of weakness over long periods of geological time depending upon the regional stress field at any given time and on the state of consolidation and stabilisation of various parts of the British crust.

Chadwick (1985) has discussed the evolution of the onshore, British Mesozoic basins in terms of the lithospheric extension ideas of McKenzie (1978). Using the deep geological data, Chadwick has shown that periods of accelerated crustal subsidence coincided with episodes of active normal faulting which probably represent pulses of lithospheric extension and thinning. The initial rift or fault subsidence caused by the onset of lithospheric thinning and the consequent rise in lithospheric isotherms was followed by a subsequent thermal relaxation of subsidence characterised by overlap beyond the rifted basin margins giving a 'steer's-head' profile. Subsidence curves and field observations from southern Britain suggest that this 'steer's-head' profile roughly occurs at the scale of the geological system, or less, as well as at the scale of the Mesozoic Era as a whole. Tectonic inversion, characteristic of onshore southern Britain in Tertiary times, is discussed by Chadwick (1985) in terms of the reverse process of lithospheric compression and thickening. In this process the thickened lithosphere causes a depression in the isotherms so that following an initial reverse faulted uplift there is a subsequent thermal relaxation giving a general uplift. In addition to providing a mechanism for inversion and its accompanying reverse faulting and folding the process, of course, would explain the associated erosion.

In conclusion, the new deep geological data base is providing 'hard' data which allows detailed interpretation of the stratigraphy, sedimentational history and structural configuration of certain areas for the first time. Much discussed geological problems, which have given rise to controversy for many years on the basis of field evidence, have been resolved or are capable of resolution by using the powerful combination of drilling, geophysical logging and seismic reflection profiling. In particular, basin areas with sedimentary cover are well understood in three dimensions for the first time allowing the possibility of further analysis in terms of palaeo-geography, distribution of facies, basin and crustal evolution. The recognition of the relationships between Tertiary tectonic inversion, Mesozoic normal growth faulting and the occurrence of 'basement' thrusts in southern England invites application to the growth faulted Upper Palaeozoic sedimentary coversequences of northern England (see Smith *et al,* 1985) with implications for the structure of the concealed Caledonide 'basement' and the unconformably overlying Permian and Mesozoic sequences. Areas with lesser economic potential than the sedimentary basins, however, where the sediment cover is thin and shallow, and regions of outcropping deformed or basement rocks, urgently require the acquisition of more deep geological data. Only when the seismic response of these areas is properly known will it be possible to assess fully their influence and behaviour as a floor to the sedimentary basins and their significance in the evolution of the continental crust beneath southern Britain.

Acknowledgements This lecture is based upon the work of many people. In particular it summarises some of the work of individual members and teams of the BGS Deep Geology Research Group to whom the writer is indebted.

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Read at the Annual Conference of the Ussher Society, January 1986

Are Jurassic sedimentary microrhythms due to orbital forcing?

M.R. HOUSE

House, M.R. 1986. Are Jurassic sedimentary microrhythms due to orbital forcing? Proceedings of the Ussher Society, 6, 299-311.

The variety of sedimentary microrhythms in the Jurassic is reviewed in order to demonstrate their occurrence in virtually all facies. Any consistent explanation must explain this variety. Many of the suggested causes are only relevant to particular cases and hence may be only proximal causes. The possibility that metronomic orbital variations resulting in climatic changes may be a controlling factor to explain all examples is argued Resultant sea-level and sedimentation factors are considered. It is argued that climatic changes may have had local biotic effects so that the units of a microrhythm can in some cases be explained largely in terms of a biotic modification of a relatively homogeneous sedimentation.

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Introduction

The pervasive presence of small-scale sedimentary rhythms of the order of 0.5-2.0m is well documented in the stratigraphic record from the Tertiary (Gilbert 1895) to the Precambrian Banded Ironstone Formations (Williams 1972). They are a fundamental property of sedimentary rocks and pose a major problem requiring interpretation.

Such rhythms are well illustrated in Jurassic sequences (Fig. 1) and particularly those of Dorset. There has been much discussion of the Lias type rhythms of limestone and marl (Hallam 1964), or rhythms of black shale, grey shale and limestone, or of black shale, grey shale of the Kimmeridge (Cox and Gallois 1979). Yet the rhythmicity pervades every Jurassic facies, and an interpretation which takes into account all the evidence and all the patterns and facies fabrics is necessary. A review of these cycles mainly for the Dorset Jurassic will indicate the range of problem requiring solution. So far interpretation has concentrated on the explanation of local phenomena, but elucidation of primary causations is also required and consideration will be given to the many types of hypothesis which have been proposed. Conclusions will be given discussing the role of orbitally forced changes.

There are rhythms at a large scale in the Jurassic (Fig. 2) which have been commented upon over the last 150years and which have been reviewed by others (Arkell 1933, Hallam 1978, 1981). Whilst some of these may reflect eustatic changes others may be due to local or regional basinal faulting or other cause. These are only of concern here in providing background changes for the micro-rhythmicity discussed.

Rhythm Patterns

Hettangian

Rhythmicity in the Dorset Blue Lias west of Lyme Regis (Fig. 3) commences immediately above the White Lias but Beds H 1 to H25 are referred to the Triassic and Beds H26 to Bed 20 to the Hettangian. Typical asymmetric cycles (or true rhythms) may commence with winnowed levels and brown to black kerogen-rich shales followed by medium grey shales and then limestones units (Fig. 4a) the tops of which are sometimes burrowed, or show *Gryphaea-,* ammonite-, or brachiopod-rich surfaces (Lang 1924, Hallam 1960, 1964, Morris 1979, Weedon 1986). Palaeoecological interpretations have been published (Sellwood 1970, Sellwood *et al.* 1970) which show the increasing evidence of oxicity through the rhythm with an increase in burrowing: the activity of burrowers may be indirectly responsible for the enhanced calcareous cement through .the sequence. In addition there are simple alternations and indications of symmetrical cycles (true cycles). No Markov Chain analyses appear to have been published. Four limestones are particularly thick (H26, 30, 42, 54) with perhaps five or six smaller limestones between them (compare with the similar indication on Fig. 1). Such indicated or inferred rhythms suggest some 22 microrhythms in the sequence of 12.34m.

Similar sequences are well documented elsewhere, for example along the Somerset coast (Wobber 1965, Whittaker and Green 1983), southern Germany and southern Italy (D'Argenio and Vallario 1967).

Sinemurian

At Lyme Regis (Fig. 3) Beds 21 to 99 are referred here but significant non-sequences are recognised in the upper

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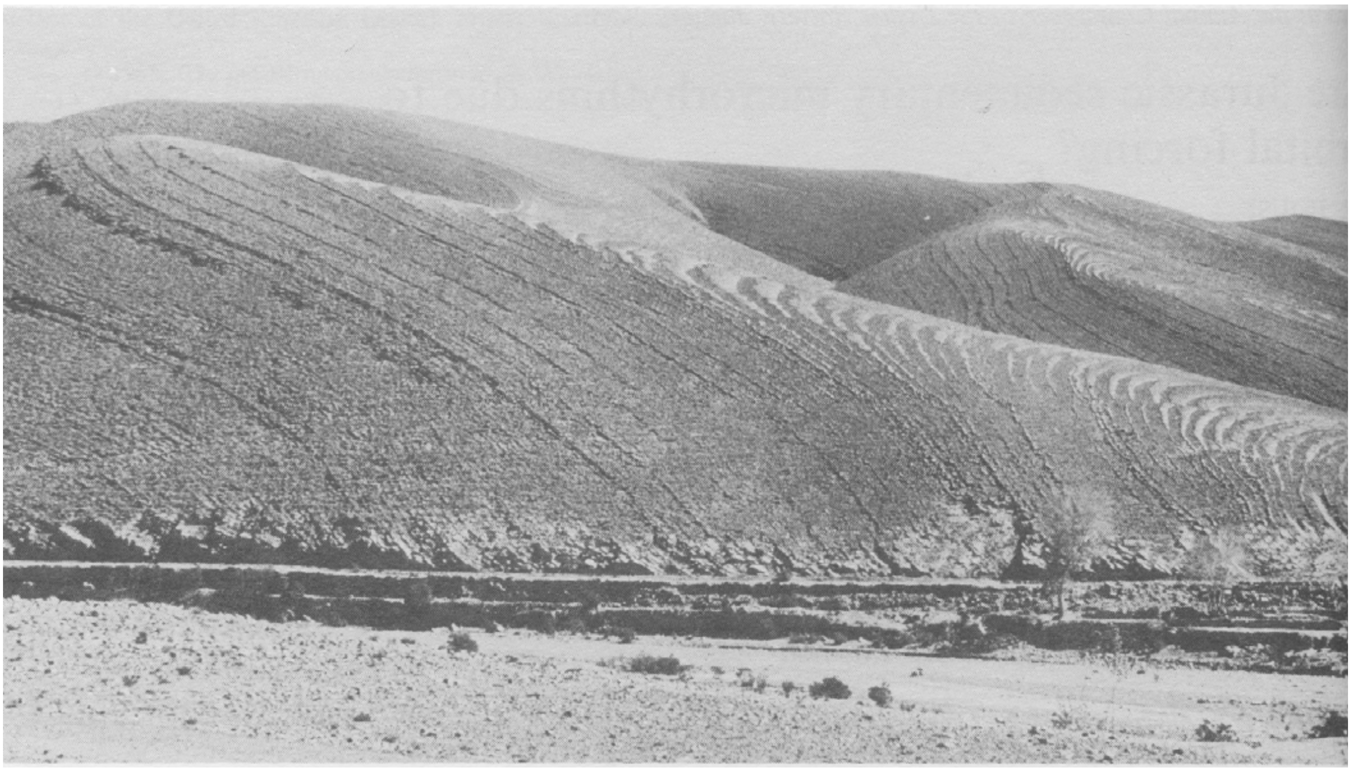

Figure 1. Jurassic sedimentary microrhythms exposed on the Meknes to Ei Rachidia road, SE of Rich, Morocco. Note the suggestion of harmonics at about every fourth or fifth microrhythm.

part. The top of the Blue Lias is taken at Bed 53 (Table Ledge) but the typical microrhythms of the Hettangian have by then been mostly lost by loss of the limestone unit in the rhythms. Between Beds 35-37, 41-43 and 45-47, an alternation of paper shales and grey marls suggest the rhythmicity continues but the limestone stage is not reached. The situation is similar to that of the Hettangian Beds H26-54 but in that case the limestone units are only reduced.

The overlying Shales with Beef (Beds *54-75,* Lang *et al.* 1923) show a marked reduction in prominence of limestones and the sequence comprises marls with many seams of fibrous calcite ('beef') which in their scale (Fig. 3) echo the earlier rhythmicity. Whilst the 'beef levels are not primary, but diagenetic (Marshall 1982), they seem likely to have developed at levels of particular primary differences. It is only in Bed 74 that kerogen-rich paper shales are important. The change in facies is usually attributed to deepening. The succeeding Black Venn Marls (Beds 76-104, Lang *et al.* 1926, Wilson *et al.* 1958) is of more monotonous dark marls and shales in which rhythmicity still has to be adequately described.

For the upper part of the Sinemurian Sellwood (1972) has given details of successions in other areas, for example in Skye, with coarsening upward cycles in silts and sandy shales. The Yorkshire coast section at Robin Hood's Bay (Units 1-JA of Sellwood 1972) shows seven rhythms in the Raricostatum Zone, Similar rhythms are seen within the Frodingham Ironstone, and especially within the overlying "Sandrock" at Roxby, South Humberside, in which Sellwood also notes seven rhythms in the Raricostatum Zone. In the Wutach area of southern Germany (Fig. 4b) in the Semicostatum Zone micro-rhythms of Blue Lias type were formed.

Pliensbachian

In Britain the earliest Pliensbachian represents a transgressive level and parts of the latest Sinemurian (Raricostatum Zone and often earlier) are represented by nonsequences. Hallam (1978) has shown how widespread this is internationally and relates it to a eustatic sea level rise.

The Lower Pliensbachian (Carixian) west of Eype Mouth comprises the Belemnite Marls (23m) with the Armatus Limestone at the base and the Belemnite Stone at the top, followed by the Green Ammonite Beds (23-34m). The Belemnite Marls (Beds 106-120, Lang *et al.* 1928, Sellwood 1972) show an alternation of argillaceous calcilutites and calcareous clays. Beds 112-119 (10.2m) show 35 alternations with some well-developed kerogen-rich paper shales. Again there is evidence of subsidiary rhythmicity, but of a different type, for example in Bed 115 (Upper Pale Band) in which Lang noted 16-20 "alternate stripes of paler and somewhat darker material". The Green Ammonite beds (Beds 122-130) show marked microrhythmicity at several levels. Bed 128 (6.86m) shows thick blue-grey clays interrupted by four thin shaly and red horizons and this is even more emphasised in the Red Band (1.7m) in which three red limestones alternate with blue-grey clays.

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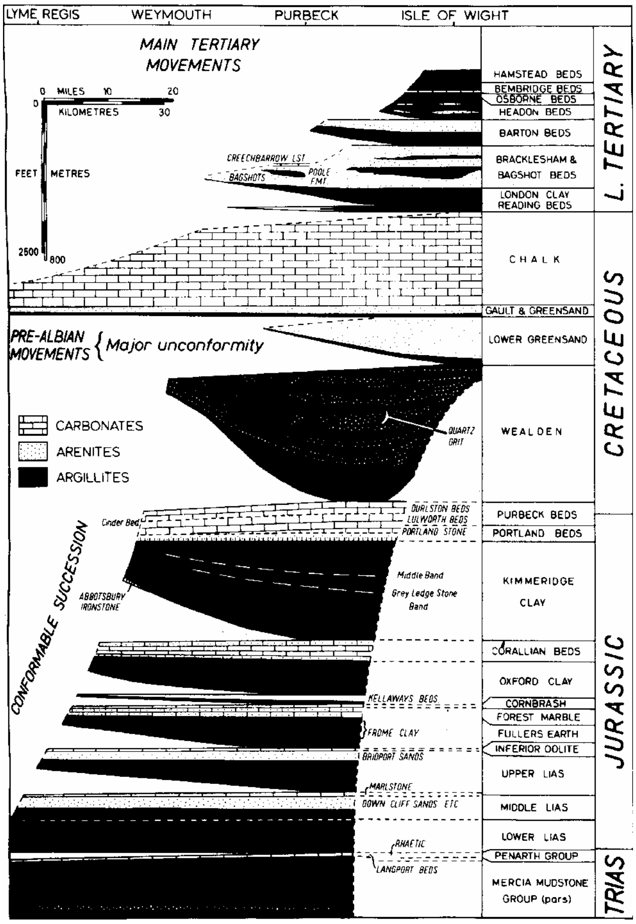

Figure 2. The Dorset geological succession drawn to emphasise major sedimentary rhythms, especially those of the Jurassic (after House 1983).

The Dorset Upper Pliensbachian (Domerian and approximately Middle Lias) is over 125m thick and the base is taken where Pleuroceras enters three metres below the top of the green Ammonite Beds and the base of the Three Tiers. A shallowing of facies is indicated through several units to the Marlstone Rock Band (0-0.6m), a condensed unit, at the top. Microrhythmicity in the Domerian is better described in the sequence of the Yorkshire coast (Howarth 1955) as alternations of calcareous sandy levels with shale, or ironstone and shale alternation (as in the Pecten Seam). Coarsening upward microrhythms have been described (Hemingway 1951, Rawson et al. 1983).

Toarcian

The Dorset coastal section commences with a very reduced limonitic oolite and micrite with algal laminations at the top of the Junction Bed which are thought to represent much of the stage (Cope, Getty et al. 1980). The Down Cliff Clay (21m) and most of the Bridport Sands (over 42m) are contained within the Levesquei Zone at the top of the stage.

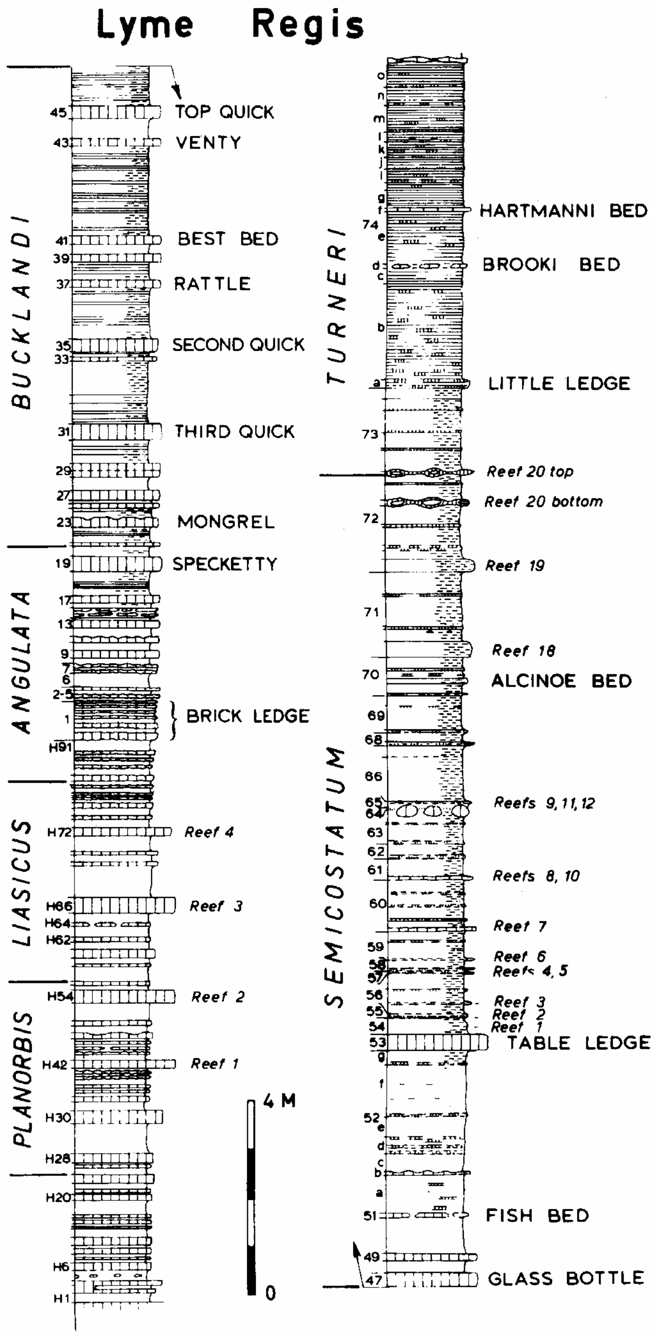

Figure 3. Succession microrhythms in the early part of the Lower Lias west of The Cobb, Lyme Regis. Based on published work of W.D. Lang (after House 1985b).

In the Yorkshire coast sequence where basal sandy Grey Shales pass into the finely laminated kerogen-rich shales of the Jet Rock with succeeding shales becoming less kerogen-rich and less calcareous indicating increasing sea floor oxicity (Hemingway 1974): at the top the Blea Wyke beds show shallower sandstones and shales so that a major shallowing upward sequence is shown by the stage. Within this major cycle detailed microrhythms have been well documented for the Grey Shales (Howarth 1973) and Jet Rock and Alum Shales (Howarth 1962). They mostly

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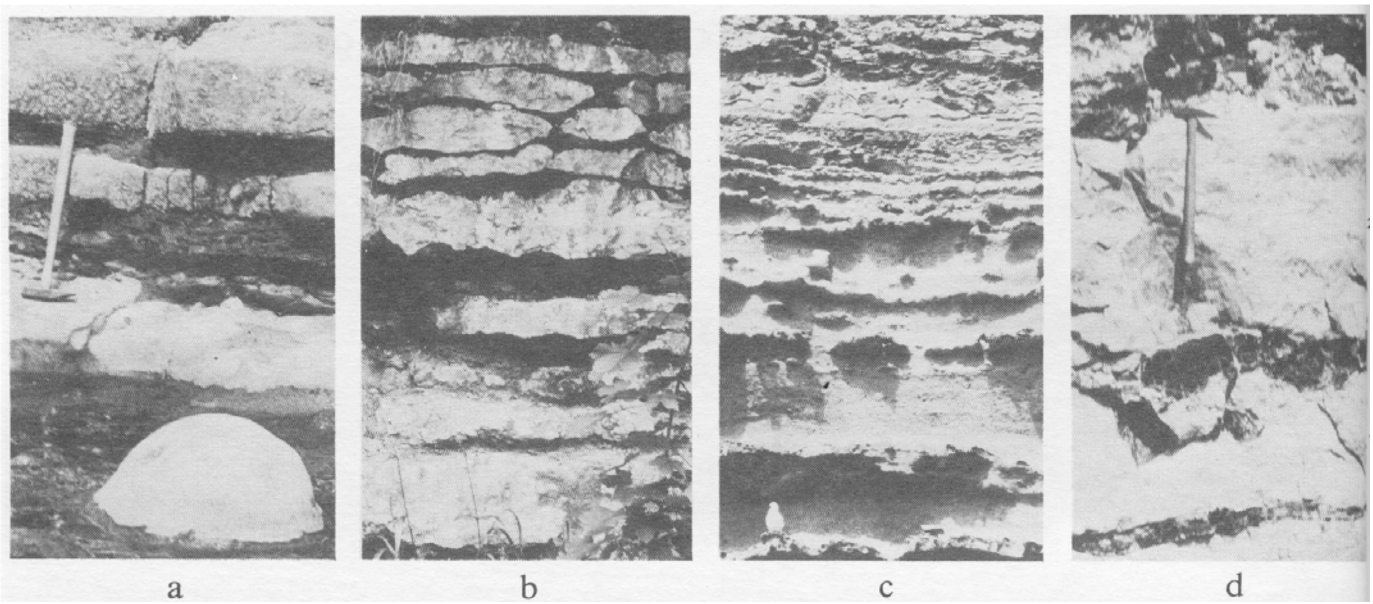

Figure 4a. Microrhythms in the Lower Lias west of The Cobb, Lyme Regis, and east of Seven Rock Point, Dorset. Showing a black shale, grey shale, and limestone triplet. b, Sequence of Lower Lias (Sinemurian) limestone/marl microrhythms in the Uberes Aubachtal, Mundelfingen, S. Germany. c, Microrhythms of friable sandstone and calcareous sandstone in the Bridport Sands (Upper Lias and earliest Inferior Oolite, Toarcian/Aalenian) at the cliff top west of Burton Beach, Dorset. A seagull gives the scale. d, Microrhythms with irregular cherts and limestones, Cherty Series, Portland Beds, Freshwater Cove, Portland, Dorset.

comprise an alternation of thin calcareous shale or calcareous nodule horizons within thicker clay successions.

For the uppermost parts of the Toarcian (much of the Levesquei Zone) the best microrhythmicity is shown by the Bridport Sands (Figs. 4c, 5) which shows friable sandstone and calcareous sandstone alternations (Davies 1967). The crumpled micas in the former, contrasting with the straight mica laths of the latter, suggest that cementation of the calcareous sandstones was early, before significant compaction which would have crumpled the micas. The calcareous, hard layers show great richness in burrows of several types, including, *Thalassinoides, Di£1ocraterion and Rhizocorallium.* In the friable sandstones depositional lamination is often preserved and there are some burrows, and *Monocraterion* has been reported. Some lower levels show cross bedding which may even cut out earlier alternations, indicating strong current activity for which a storm depositional environment may be appropriate, but this is not the rule. The calcareous levels may represent either stillstands when burrowing was favoured or suitable climatic conditions. It would follow that, since active burrowing would lead to partial pressure changes following upon respiration activites of the burrower, carbonate cementation is likely at such times. Indeed, the correlation between burrowing and cementation is so great in the Jurassic as to suggest a casual relationship as suggested for the Blue Lias pattern.

Aalenian

The upper two metres of the Bridport Sand at Burton and the lower Inferior Oolite are referred to the Aalenian. Figure 4c illustrates the transition and shows how micro-rhythmic units progressively thin. The Lower Inferior Oolite is only represented at Burton by the Scissum Beds (1.3m) and the whole suggests an increasingly shallowing environment. Rhythmicity is better shown in France (see below).

Bajocian

The Dorset Inferior Oolite is a condensed deposit (up to 6m) with discontinuities in which a record of microrhythms is not clear. In Yorkshire the essentially alluvial sequence of the Deltaic Beds is interrupted by the marine pulses of the Eller Beck, Blowgill, Whitewell. Yons Nab and Scarborough units of the Lower Bajocian, but these correspond to changes at a larger scale than the microrhythms under consideration. But the Lincolnshire Limestone (probably Discites and Laeviuscula Zones) is interrupted by bedding planes which suggest the microrhythmicity. Certain of these are early lithification hardgrounds with adnate oysters and serpulids and often with burrows or borings which can be traced for some distances (Marshall and Ashton 1980).

For the Upper Bajocian it is necessary to look abroad for evidence that microrhythmicity continues in appropriate facies. In the Digne area, SE France, superb sequences of alternations of marl and limestone compare in excellence with those of the Dorset Blue Lias (Pavia 1971). At the Revin de la Couete Chaudon, for example, in the Subfurcatum Zone, 42 such units are developed in 21m in parts with so strongly a metronomic regularity that estimates could be made where there are gaps without

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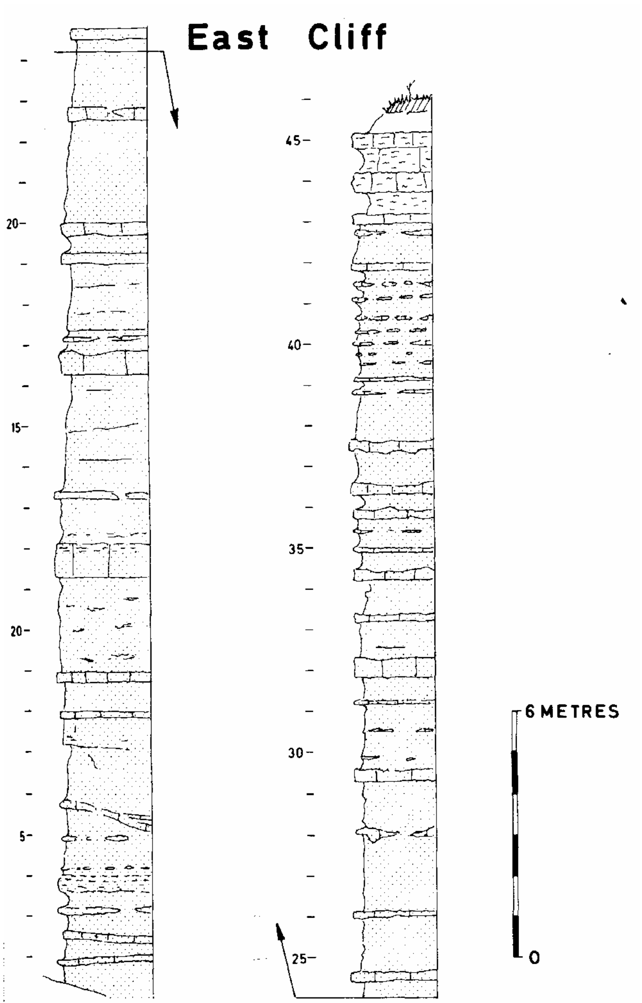

Figure 5. Succession of microrhythms in the Bridport Sands and overlying Inferior Oolite at East Cliff, West Bay.

limestones (following the crude method suggested by House 1985b).

Bathonian
The sequence in Dorset comprises the Fuller's Earth, Frome Clay, Forest Marble and Lower Cornbrash only the last of which is adequately described (Arkell 1947). The best study of the earlier levels is of the Horscombe Vale No. 15 Borehole in the Bath-Frome area (Penn et al. 1979) where the Fuller's Earth is 46m thick. Three rhythms are recognised in the Lower Fuller's Earth which typically commence with a shell bed of epifaunal forms followed by calcareous mudstones dominated by infauna, and followed by silty mudstones dominated by nekto-planktonic forms, such as Bositra, and the rhythms are terminated by a thin, often porcellanous limestone with a burrowed top surface. Above the Fuller's Earth Rock the Upper Fuller's Earth Clay shows five cycles (perhaps more) in which an additional black shaly mudstone phase is added. These rhythms are interpreted as deepening upward rhythms with a sudden return to shallow water at the top that is, in terms of rapid epeirogenic uplift and gradual epeirogenic downwrap (Penn et al. 1979). Within these rhythms in the borehole are microrhythms of limestone and mudstone at a finer scale (Units 10-13, 19-24). In the East Midlands rhythmicity is shown in the Upper Estuarine Series (Aslin 1968). Rhythms are also described for the Hampen Marly and White Limestone Formations (Palmer 1968).

Callovian
The initial deepening is shown in South Dorset in the transition from limestones of the Upper Cornbrash (5m), through the Kellaways Clay (2.5m) to the sandy doggers of the Kellaways Rock (0.7m) but this is poorly exposed at present. The second deepening phase is seen in the bituminous shales of the succeeding Jason and Coronatum Zones. These are followed by clays of the Athleta Zone and in them layers of concretions and septaria occur but not details of microrhythms have been published.

For this interval the best account of microrhythms is that of Brinkmann (1929) at Peterborough who distinguished two types of microrhythm (Arkell 1933). The 'roofed' type (Dachbank) commenced with green then brown days, passed up into bioturbated levels with comminuted and winnowed shells above and an oyster plaster at the top which was interpreted as a pause in sedimentation. By contrast in the 'floored' type (Sohlbank) a pavement of crushed ammonites forms the sole or base which is followed by an epifauna shell bed, then Nucula-rich clays and then a thin breccia of "Pseudomelania" after which a pause in sedimentation is followed by another ammonite pavement. The preponderance of nekton and plankton in the shales has led to the interpretation that the sea floor was anoxic when they were formed whereas at the times of formation of epifaunal shell beds the sea floor was oxic but this has been debated (Duff 1975, Martill 1985). Callomon (1955) was able to match many of Brinkmann's levels near Oxford.

The uppermost 6m of the Callovian, including the Lamberti Limestone, was studied by Hudson and Palframan (1969) at Woodham, Bucks, without any reference to microrhythmicity, and they recorded epifauna in all their beds (0-11). Some six 'hard' beds were noted which suggests that a microrhythmicity similar to that of the earlier Callovian was continuing but without anoxic sea floor environments; pyrite might form, nevertheless, within the sediments.

Oxfordian
East of Weymouth 12m of clays of the Mariae Zone and 12m of clays of the Cordatum Zone (preceding the Nothe Grit) are exposed in Furzy Cliff (Cope, Duff, et al. 1981). Microrhythmicity has not been described but it shows as plasters of Lopha gregarea and as red marl seams and red nodule seams in the upper part. At Woodham

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rhythmicity was not noted in the early Mariae Zone (Hudson and Palframan 1969) although one 'hard' bed was described above the Lamberti Limestone. Rhythmicity at rather less than one metre intervals is seen as pale bands within grey shale in the Mariae Zone NE of Scarborough Castle.

Microrhythmicity is better documented for the Oxfordian by Cariou (!966) from Poitou as rhythms of one to three metres in thickness which commence with fossiliferous grey marls with rounded clasts and winnowed shells, followed by marls, often concretionary, and terminated by thin red or grey limestones with bored upper surfaces.

The sedimentary rhythms of the Dorset Corallian have been well described and much debated (Arkell 1933, 1947, Talbot 1971, Wilson 1980). These are of a different facies to those already described and, if the ammonite zones are any guide to time, are of a different order (as may be those discussed under Bathonian). The four rhythms have been interpreted as representing environmental changes from offshore shelf to near sub-tidal or even lagoonal on the interpretation of Talbot who prefers not to follow Arkell and take the clays as initiating upward-shallowing rhythms (apart from the first starting with the Oxford Clay) and assumes that the Hudlestoni Bed (Preston Grit), Qualiscosta Bed and Clavellata Beds represent the deepest phase. I do not agree with this since this facies, with large bivalves, is similar to that seen today following particularly heavy storms around Torbay where large articulated shells including the great Torbay Cockle can be cast up into even a supratidal environment. A transgression following such facies would give the proximal pebble beds as a prelude to significant deepening represented by the clays. Following Klupfel (1926) all authors base their interpretation of these rhythms on relative sea-level changes. Even within these rhythms, however, at many levels a microrhythmicity is common but, given the changing facies, impossible to document for any great thickness.

Kimmeridgian

This term is used in the traditional Dorset sense (following Cope, Duff *et al.* 1980) but international Use would restrict it to the Lower Kimmeridge Clay up to Blake's Bed 42 (Fig. 5) and the top Jurassic, above that level, would be termed Tithonian or Volgian.

Brookfield (1979) has argued that the deepening initiating the Kimmeridge Clay is more accurately drawn in Dorset at the base of the Mutabilis Zone and he uses Blake's term Passage Beds for units from that level down to the top of the Sandsfoot Grit. Microrhythmicity described by Brookfield is best developed in the Sandy Clays of the Cymodoce Zone where they comprise a basal lag horizon of phosphatic pellets and shells resting on an erosion surface, followed by fining upward clayey sand passing into blue clays which, when not eroded, show *Thalassinoides* burrows at the top. These he interprets as being produced by storms on a shelf, or perhaps as distal barrier bar inlet deposits, presumably also storm related. Cox and Gallois (1981) describe such rhythms as their Type A. Lower Kimmeridge Clay fabrics have been described by Aigner (1980).

Microrhythmicity is perhaps better seen in equivalent levels described by Enay (1980) who has traced individual rhythms from Normandy to SE France. These show sequences of marl to limestone with bored upper surfaces

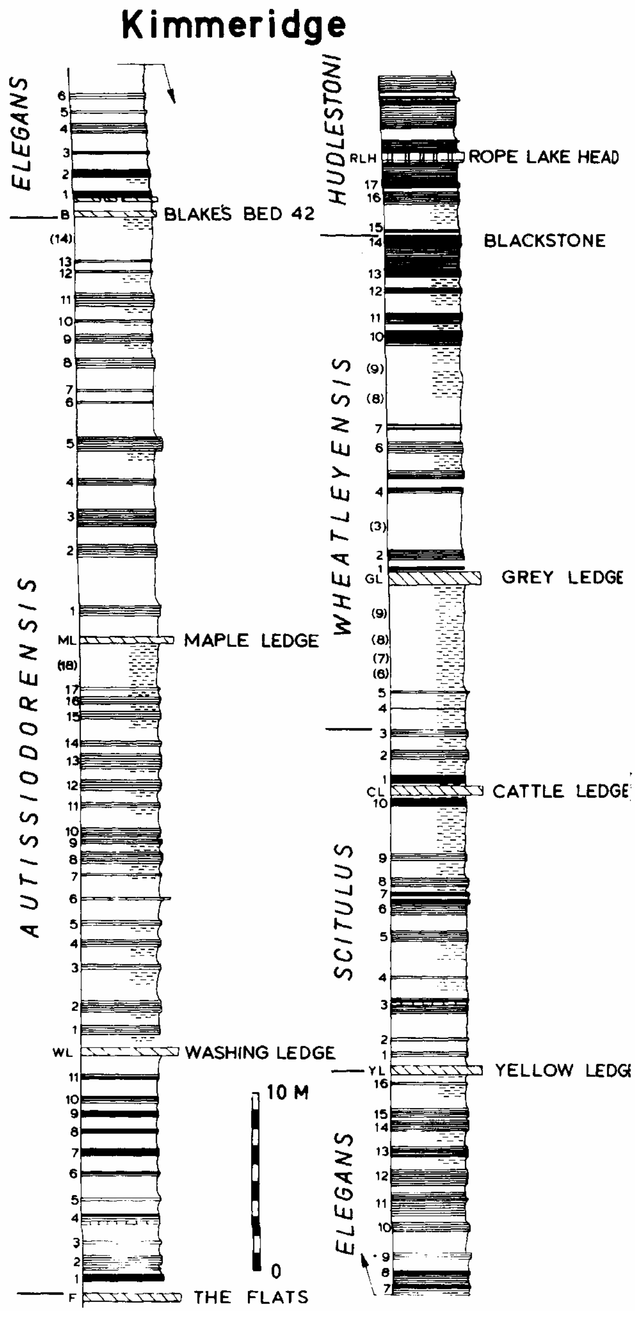

Figure 6. Succession of microrhythms in the Kimmeridge Clay around Kimmeridge from west of Gaulter Gap (below) to Cavell's Hard (above). Based on published work of B.M. Co and R.W. Gallois (1979) (after House 1985).

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and Enay interprets these as shallowing upward sequences related to land emergence in agreement with the views of Cariou (1966).

It is in the top Lower and early Upper Kimmeridge Clay around Kimmeridge (Fig. 6) that microrhythmicity has been best described, firstly by Downie (1955, and in House 1969) who recognised both rhythmic, alternating and cyclic arrangements, the last showing a sequence of clay, kerogen-rich shale, kerogen-rich marl, coccolith limestone, kerogen-rich marl etc. The most detailed account is by Cox and Gallois (1981) who consider a basic microrhythm to comprise a basal fissile oil-shale with phosphatic debris passing up into pale and very pale mudstones commonly with limestone doggers or septaria (their Type B). In the Autissiodorensis and Wheatleyensis Zones, allowing for gaps, some 98 micro-rhythms have been estimated (House 1985b).

Contrasting *interpretations* of the microrhythms have polarised into one model emphasising water column stratification and occasional sea floor anoxia (Tyson *et al.* 1979) or a model dependant on variations in the plankton (Gallois 1976). The substantial more recent geochemical work (Williams *et al.* 1981, 1983) has not rigorously analysed differences precisely through micro-rhythms and the results are correspondingly generalised. A Fourier analysis by Dunn (1974), using geochemical data, claimed evidence of multiple cycles and evidence of orbital forcing.

Rhythmicity through the upper part of the Upper Kimmeridge Clay has not been documented in detail and only occasionally shows evidence of changes at a micro-rhythmic scale. At a larger scale Wimbledon (1987) has recognised faunally related cycles approximating to the ammonite zone and subzones.

Portlandian
The classic account of the Portland Beds of the Dorset mainland (Arkell 1935, 1947) shows irregular micro-rhythmicity throughout the sequence as 'hard' bed, marly clay alternations in the Portland Sand, limestone/chert alternations in the Cherty Series, and periodic bedding interruptions throughout the sequence (Fig. 7). The limestone/chert alternations (Fig. 4d) have been explained by Townson (1975) as due to alternations in the percentages of the sponge *Rhaxella* which seems to have been the source of the chert. Bedding plane interruptions in limestone sequences seem part of the microrhythmicity but are less readily generalised. Again rhythmicity at a larger scale corresponding to ammonite zones and subzones has been documented by Wimbledon (1987) and, since these introduce novel faunas, an explanation in regional deepening seems reasonable.

Purbeckian
The Dorset Purbeck Beds and Wealden Beds are essentially non-marine but in them many ostracod cycles have been described (Anderson and Bazley 1971, Anderson 1985) which have been widely traced in southern England. These show an alternation of marine and non-marine types which Anderson ascribed to climatic cycles (Sandy 1985). Here the obvious contrast is faunal rather than lithological.

Types of explanation
A wide range of hypotheses has been put forward to explain small scale rhythmicity, both for the Jurassic and for other periods. For the convenience of a brief review, these may be classified into causations which are locally produced, globally caused, or orbitally forced (Fig. 8). There can be little doubt that all the illustrated causes, and others, have some effect in controlling the sedimentary record of microrhythmicity at certain times and places. The problem is to elucidate which may be the greatest control, or rather which, if any, may be regarded as the primary causation, and which may best explain the

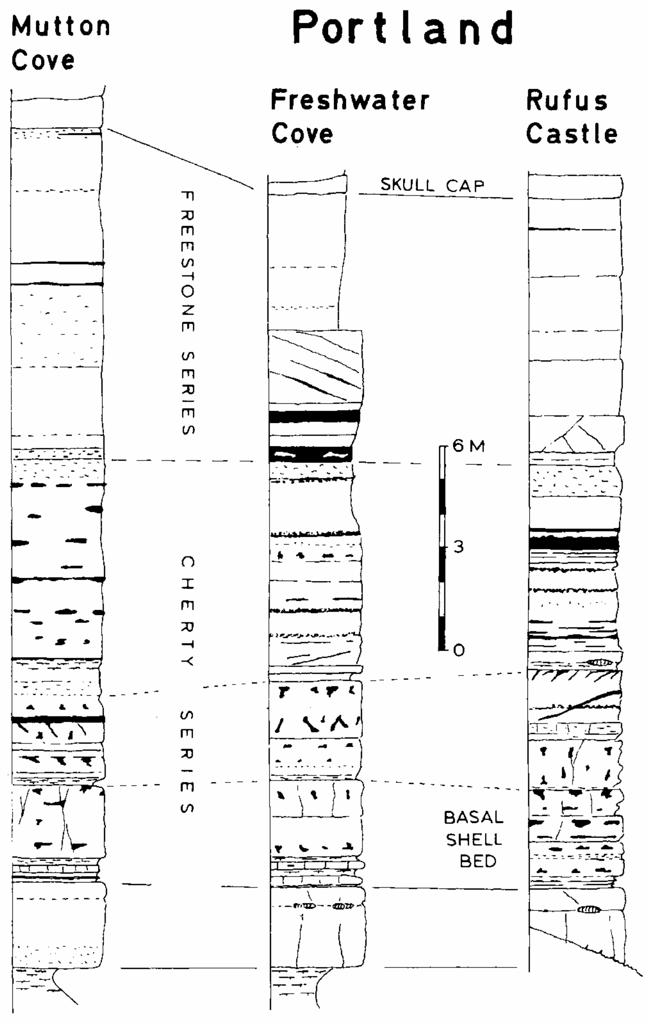

Figure 7. Microrhythms and bedded units in the Portland Beds of the Isle of Portland.

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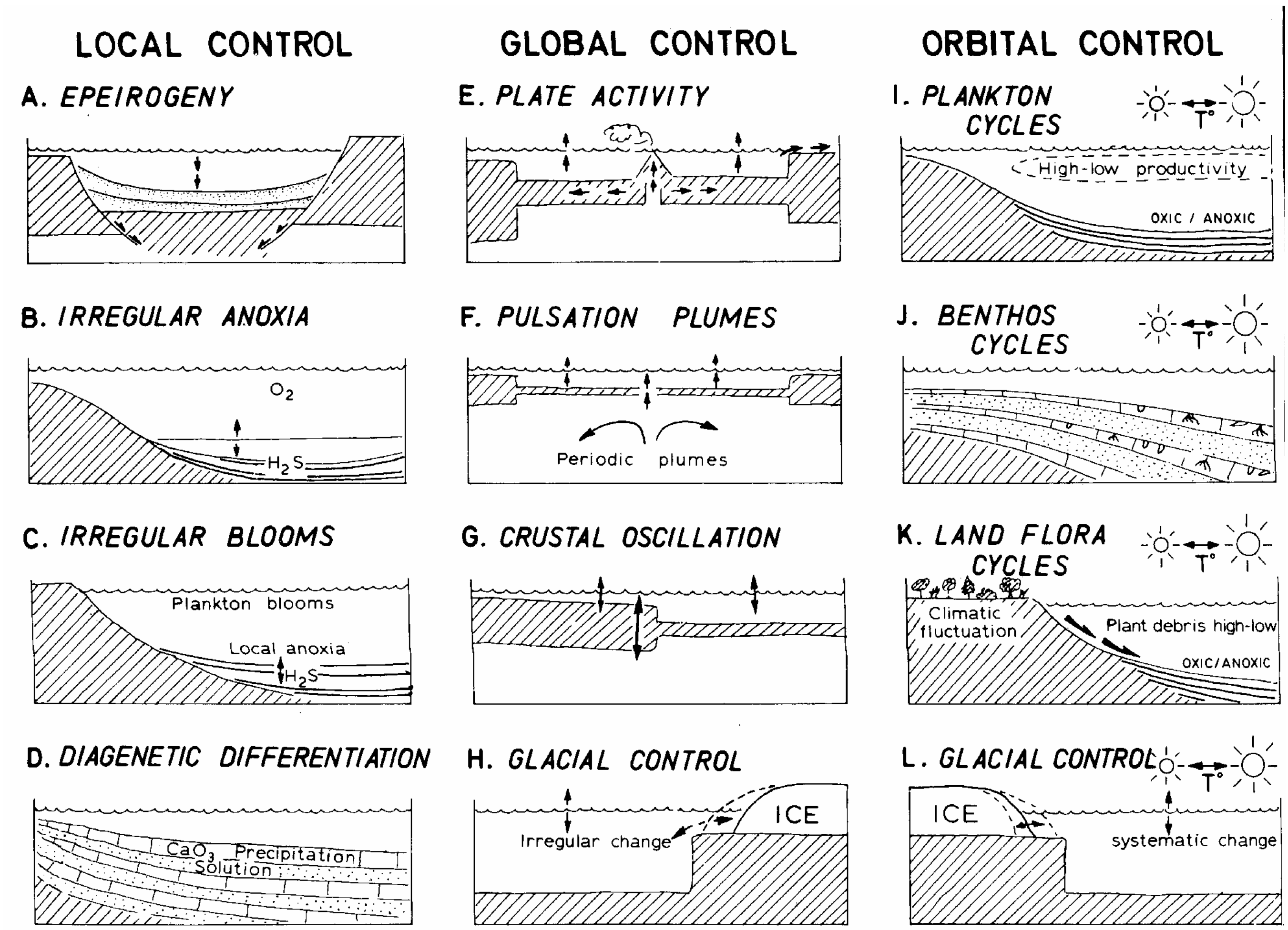

Figure 8. Cartoon illustrating some of the many hypothoseses put forward to explain sedimentary rhythmicity.

pervasiveness of the phenomena in a range of facies, through much of the stratigraphic record, and at a scale which varies little, either in thickness (0.2-2m) or in estimated duration.

Local Controls

Local fault controls of basin subsidence is the commonest type of general hypothesis invoked (Fig. 8A) and this links causation with sea-level and erosion cycle changes resultant upon local basin faulting. This may explain some of the larger-scale phenomena (Fig. 1). On this model the regular small-scale rhythmicity would be caused by the intermittent release of a progressive stress build-up due to large-scale tectonic controls.

Such hypotheses were considered by Klüpfel (1916, 1924), Frebold (1927) and others in detail for the German Jurassic in relation both to sea-level changes and changes in sediment supply following faulting. Bayer, Altheimer and Deutschle (1985), Guidish *et al.* (1985) and Watts (1982) have considered this more recently. This type of model has applicability in most of the facies regimes discussed above but it does not explain readily the occurrence of expected shallowing upward as well as deepening upward rhythms or even symmetrical cycles, nor adequately resolve the widespread similarity in microrhythm scale. It is, however, bound to be a factor important to take into account and the European Jurassic must be seen in the context of continental separation (Heller and Angerine 1985).

Other local control hypotheses include sea-floor anoxic changes (Fig. 8B) in which movements of an O_2/H_2S boundary (Tyson *et al.* 1979) would control sedimentary types such as Kimmeridge Clay black shales. Other interpretations would prefer the influence of periodic plankton blooms (Gallois 1976) (Fig. 8C). Aigner (1982) has claimed episodic, storm-dominated controls for Muschelkalk microrhythmicity, but a primary cause for periodic storms seems more likely. The control of diagenetic changes in producing the rocks now seen is obvious, but diagenetic change (Fig. 8D) has been invoked as the only, or at least, the main cause of rhythmicity in limestone-shale rhythms (reviewed by Einsele 1982). Brongsma-Sanders (1971) has suggested how climatic change might initiate cyclicity in evaporite and bituminous rocks and variations of this have been used to explain patterns in some ocean cores (Habib 1983, Gradstein and Sheridan I983). These are all local solutions which may, or may not, be correct as to

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proximal cause but are applicable to limited facies only and which do not address the problem of primary causation.

Global Controls

There are a number of mechanisms which might give sea-level changes which would be eustatic (Fig. 9) and which could embrace the controls invoked for local models. Such mechanisms include plate margin activity (Fig. 8E), which may be either volume changes at constructional margins or tectonic factors at constructional margins. Donovan and Jones (1979) argued that changes in the ocean ridge systems are long term events (10-30 Ma) and this might make them not applicable to microcycles, but oceanic volcanicity, given periodicity if related to release of a regional plate stress system, might be a possible cause for small-scale effects. Hallam (1969, 1978, 1981) has considered the major Jurassic cycles (Fig. 1) and claims that some of these are related to eustatic cycles, but they are of a substantially greater scale than the microcycles which is the concern here. Mechanisms of these sorts, however, would essentially give relative sea-level rises and leave the same problems of rhythm pattern, scale and duration unresolved as for the local tectonic control hypotheses outlined above.

Other possible models include 'pulsation tectonics' (Fig. 8F) resulting from periodic mantle convection plumes (Sheridan 1983) but the order of cycles here (ca. 65 ma) is too great for microcycles. An older generation considered oscillatory crustal movements (Fig. 8G), either in terms of large positive areas and sinking troughs (Haug 1900) or of a continental mass, but it is difficult to envisage these acting with the frequency needed to explain microcyles.

For times which have ice caps, climatic fluctuations provide an excellent mechanism, much discussed (Fig. 8H), but periodicity in this is more readily invoked by a

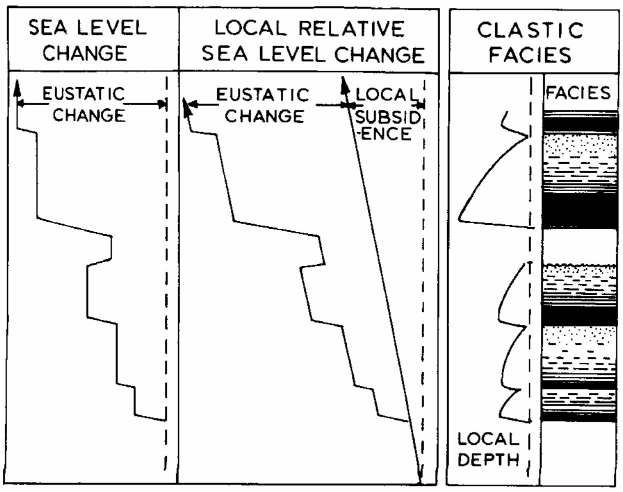

Figure 9. Illustrating the typical model for explaining micro-rhythms in terms of sea level change.

primary orbital forcing. But times, including the Jurassic (Hallam 1985), with no ice caps raise problems since in such conditions any mountain ice is likely to be volumentrically trivial: Such a control provides a mechanism for variously asymmetrical microcycles not readily achieved by other methods,

Orbital Controls

Essentially climatically controlled environmental changes which are orbitally forced by modifications of insolation received from the sun (Fig. 10) have been considered as a likely causation at least since Croll (1875) developed such views in relation to the ice ages and Gilbert (1895) invoked them for Tertiary microcycles. At the present day the main controlling cycles are the Eccentricity of the Orbit (two nutations at 106 ka and 410 ka), the Obliquity of the Ecliptic or Tilt (41 ka), Precession of the Equinoxes (23 ka) and Time of Perihelion (19 ka). Estimates of the period of these cycles vary considerably, and they would act mostly in different ways (contributions to Berger *et al.* (eds), 1984) and the interpenetration effects can give complex results (House 1985b).

With the advent of analysis of deep sea ocean cores for the last 800,000 years by Fourier Analysis, the presence of strong signatures at approximately the expected periods has resulted (Fig. 10) (Hays, Imbrie and Shackleton 1976, Imbrie and Imbrie 1980, Imbrie 1985). Thus the possibility of orbital control has now become the probability that it is an important factor in the environmental control of past sedimentation.

Anderson and Kirkland (1960) and Anderson (1984) have shown how varves in the evaporitic sequence of the Jurassic Todilto Formation of New Mexico indicate the presence of orbitally forced cycles. Evidence for the late Triassic is similar (Olson 1984). Studies of Fourier Analysis in the Kimmeridge Clay (Dunn 1974) suggested the possibility and Weedon (1986), using Walsh power spectra, has argued for cycles at 21 and 41 ka (or perhaps 21 and 31 ka) for the British Lias. On the basis of analysis of dominant rhythmic units in the Lower Lias and Kimmeridge Clay House (1985b) argued that the dominant rhythmicity could range from 36.9 to 48 ka dependant on which of the many estimates for the duration of the Jurassic (from 60-75 Ma) is used to determine the length of zones but that a 40ka seemed about right. He concluded it might be more rational to assume the dominant cycle was that of the Obliquity of the Ecliptic, and suggested that microrhythmicity was probably a better means of assessing zonal duration than the present assumption that all zones represent equal time.

Ways in which insolation controls would affect climate are complex and beyond the scope of this contribution. How these might operate locally only will be considered. Four types of proximal control on sedimentary environment are emphasised here, sea level change,

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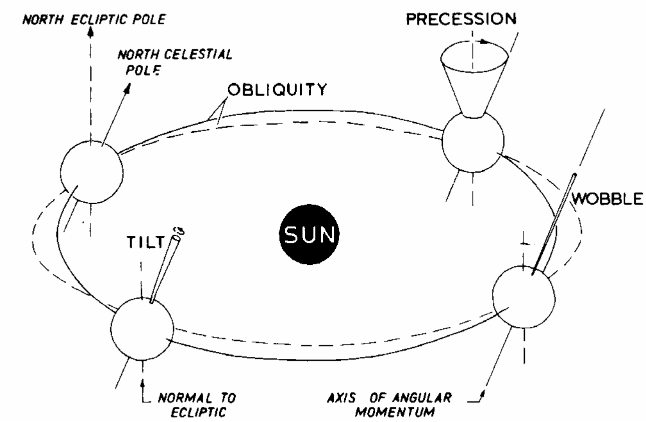

Figure 10. Diagram of the earth's planetary motion in the solar system to illustrate the several orbital characters which give cyclicity to the insolation received on the earth from the sun.

chemical change, sediment input change and biotic change. Since the first three have never lacked proponents, the last will be developed rather more fully.

The present day oceans have a volume of 1370 x 10^6km^3, an average depth of 3.8 km and-an average temperature which may approximate to 2°C. Density change of water with temperature is not regular and in fresh water below 4°C volume increases with temperature fall until freezing. For purposes of calculation if one takes a density change figure of +l x 10^{-4} per degree rise then this suggests that a sea level rise of one decimetre might occur if sea temperatures could rise by an average of one degree. But substantially greater changes are possible if surface waters, rise substantially, or if the average ocean temperature were to be significantly higher than at present. So a helpful contribution may be made to explain the many evidences of increased current-flow activity associated with parts of some microrhythms since this would follow slight deepening. Further, various asymmetrical and other patterns are easily assimilated into the model. Given ice caps and mountain ice more

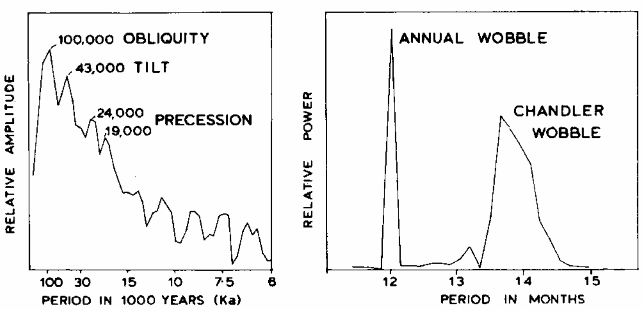

Figure 11. Left, Fourier analysis indicating peaks of relative amplitude in oxygen isotope records from the ocean cores for the last 800Ka and the interpretation in terms of orbital forcing (modified from Covey). Right, an analysis of the relative power of the Annual Wobble of the earth and the Chandler Wobble (modified from Chinnery).

spectacular effects are to be expected, for ice caps can collapse catastrophically in certain circumstances.

There are several rhythmic patterns which are more readily interpreted by the effect of multiple cycles than other models allow. A common four- or five-fold pattern in parts of the Lias mentioned above (Fig. 3), and seen in the Jurassic elsewhere (Fig. 1) can be simply explained in this way (House 1985b). So too can the declining rhythm effects illustrated by Aigner (1982) and the periodic coccolith and dolomitic levels in the Kimmeridge Clay (House 1985b).

Temperature changes of sea water would lead to changes in a wide range of chemical processes and the physical controls of local environment, sedimentation and diagenesis.

Major changes in sediment input would result from changes in base level resultant upon sea level alterations and climatic changes in bordering land areas. This would produce the local causes required by many hypotheses in terms of clay grade to sand grade sediments or even plant debris variations (Habib 1983) and storm events (Aigner 1982).

It is the faunal and floral changes dependant upon climatic changes which are stressed here because sedimentologists, almost universally, have neglected their role. Even given a steady sediment input it is possible to model microcycles by biotic fluctuation controlled by temperature alone. For example in situations of clay grade sedimentary input, firstly planktonic blooms with an anoxic sea floor resulting might give kerogen-rich shales and laminar clay particle settlement, secondly, if there was a temperature controlled change in planktonic and nektonic organisms this could lead to the production of bioflocculated clays and lessened anoxia in which certain benthonic forms could flourish changing the nature of sedimented clays: if, thirdly, the trend gave more active benthonic activity in which widespread burrowing occurred then this could lead to enhanced carbonate cements and impure limestone formation. The result might be a black shale, grey shale, and limestone of the Lias rhythm type. Note that the rate of thickness of rock accumulated per unit of time will differ markedly during the course of the rhythm. On such a model, Bridport Sand rhythms, in constant sand-grade facies, would become explicable if at a certain temperature enhanced burrowing becomes possible and with it, enhanced contemporary carbonate cementation. Portland-type rich and poor Rhaxella cycles and carbonate rhythms can be incorporated in the model with little problem. Those of the Purbeck might require sea level changes in addition to environmental evaporation controls. The major control on sedimentation which organisms can have in shallow water facies such as those of the Jurassic is fundamental. It is on biologically-controlled sedimentation that diagenetic processes work.

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However, in practice, not only will all four of the above temperature-controlled factors be important and interactive, but local and global tectonic and eustatic effects will play a role in obscuring the simple orbitally forced pattern.

In a previous paper on the subject of Jurassic sedimentary microrhythms (House 1985b) emphasis was placed on the role of periodicity in contributing towards an improved time scale. All Work on sedimentary rocks is handicapped by the inadequacies of the radiometric scale. That still is the major way in which orbital forcing analysis might enable quantification of environmental changes. In this paper the purpose has been to show how any interpretation of microcycles must acknowledge their widespread distribution in all types of sedimentary facies. This effectively eliminates a large number of hypotheses as candidates for a primary causation. If it is true, as argued here and by others, that the range of microcycles are under orbital forcing controls, and if an improved time scale can enable these to be integrated into an analysis of orbital cycles through time, then, not only will a wholly new time Scale be provided, but a revolutionary precision will be afforded to studies of palaeoecology, palaeontology, sedimentology and basin analysis.

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Read at the Annual Conference of the Ussher Society, January 1986

A high diversity ostracod fauna of late Pliocene age from St. Erth, south Cornwall

C. MAYBURY
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Maybury, C. and Whatley, R. 1986. A high diversity ostracod fauna of late Pliocene age from St. Erth, south Cornwall. *Proceeaings of the Ussher Society,* 6; 312-317.

The late Pliocene ostracod fauna from St. Erth (south Cornwall) with 378 species (all benthonic) is by far the most diverse ever described from a single deposit. Comparison between this fauna and Recent and fossil faunas around the world reveal its unique nature. The only fauna known of similarly high diversity is recorded by the authors from deposits of equivalent age in North West France. The St. Erth fauna contains many elements which do not occur in U,K. waters today and the overall aspect of the fauna is one of warm water origin. The high diversity is thought to be due to a combination of factors such as: a possible mixing of warm and slightly cooler water species; some degree of salinity variation; the wide variety of niches available in the general region and considerable allopatric speciafion. Favourable preservation has undoubtedly enhanced the diversity.

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Introduction

The late Pliocene marine deposits east of St. Erth (SW 551 351) have yielded the most diverse benthonic ostracod fauna yet described. From 23 sediment samples a total of 378 species/subspecies belonging to 77 genera/subgenera and 18 families were obtained. This level of diversity is unparalleled by any previously described fossil or Recent ostracod fauna known to the authors. Another measure of the extraordinary nature of these deposits is that from an average washed residue sample weight of 44 grams an average number of 130 species could be obtained. In absolute figures, for example, a single 50 gram washed residue sample yielded 236 species and a 6 gram washed residue sample 202 species. The incidence of the fauna is also high with 66 specimens in a single gram of washed residue. This paper describes the nature of the St. Erth deposits, the palaeo-environment in which they were laid down and its faunal composition. It demonstrates the uniqueness of the ostracod fauna by comparison with various Neogene to Recent faunas and discusses the causes of its high diversity.

Stratigraphy and Palaeoenvironment

The St. Erth beds are situated approximately 100 feet above sea level and rest unconformably on an eroded Devonian surface. The deposits themselves conform in type to a classical marine transgressive sequence (see Wilson, 1975) as shown in Figure 1. The sand layer at the base of the section varies in thickness from about lm to about 6m and exhibits a fining upward sequence; the finer deposits consist of either rule dune sand or coarser beach sand, the upper levels usually consist of beach sand, while the uppermost layers of beach sand contain a quantity of clay. It is assumed (see Mitchell *et al.,* 1973) that the sand deposit was laid down in shallow water lagoons. The sand layer is capped by a fine band of small pebbles and is

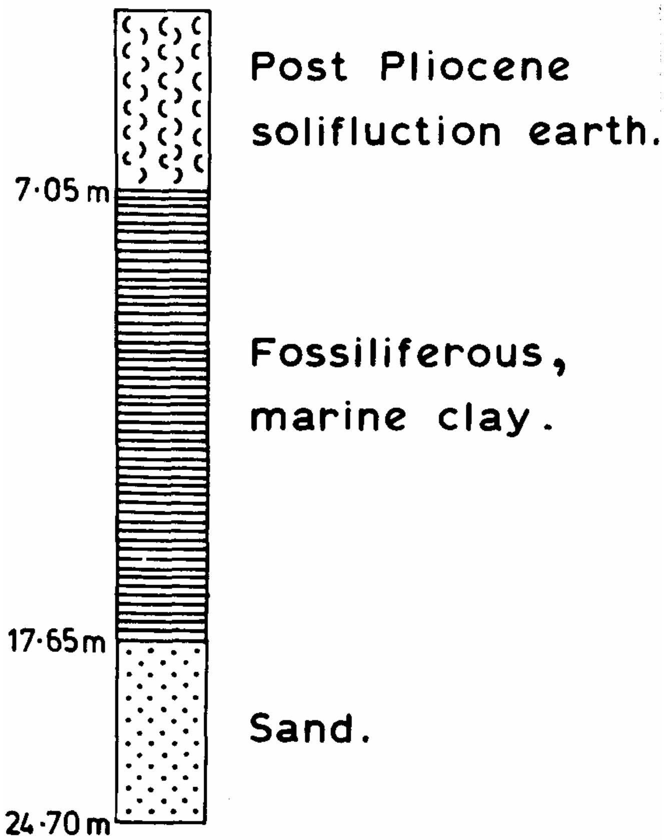

Figure 1. Generalised Section of the St. Erth Beds (after Mitchell, 1965 and Mitchell *et al* 1973)

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succeeded by highly fossiliferous marine clays. It is thought (Mitchell *et al., op. cit.)* that these clays were deposited daring a short time interval and were originally very uniform. Their faunal content includes an abundance of Mollusca, Ostracoda and Foraminifera and to a lesser extent Porifera, Coelenterata, Bryozoa, Annelida, Crustacea (other than Ostracoda), Echinodermata, Tunicata and Pisces. The floral element of the clay is inconsiderable: macrofossils include the remains of coniferous leaves, heather, mosses and liverworts and the microscopic remains comprise 58% tree pollen, 21% Ericales, 17% Chenopodiaceae, 2% Gramineae, 1% Cyperaceae and 1% unidentifiable organic material (Mitchell *et al., op. cit.).* It is impossible to state how much of the clay layer has been removed subsequent to its deposition but it is certain that the post Pliocene build up of solifluction earth is responsible for much of its preservation.

The 23 samples examined by the authors were from the clay levels of the sequence. The sand samples examined were barren of Ostracoda.

The palaeogeography and palaeoecology of the St. Erth Bed is reviewed in Maybury and Whatley (1980), Jenkins (1982) and Jenkins, Whittaker and Carlton (in press). The conclusions resulting from these studies are as follows:

1. Britain was connected to Europe in the Upper Pliocene. 2. The St. Erth deposits were laid down by a *"warm"* sea with temperature ranges similar to those of the present Mediterranean. Evidence provided by the planktonic Foraminifera (Jenkins, 1982) suggests that the palaeo-temperature was between 10 and 18°C (and probably in the upper part of this range).

3. The St. Erth clays were deposited in shallow water, possibly in a narrow gulf or strait (see Bell, 1898). The depth range was littoral to sublittoral and the maximum depth not much below low water at spring tides (McMillan in Mitchel *et al., op. cit.).*

4. Normal marine salinities prevailed during the deposition of the St. Erth clays, although there is evidence that the St. Erth sands may have been redeposited in muddy, probably transitory, coastal lagoons (Catt and Wier *In:* Mitchell *et al., op. cit.).* Also the presence of a high number *of Leptocythere* species at St. Erth may be indicative of a brackish water influence obtaining during some stage or stages of deposition (Maybury and Whatley, 1980).

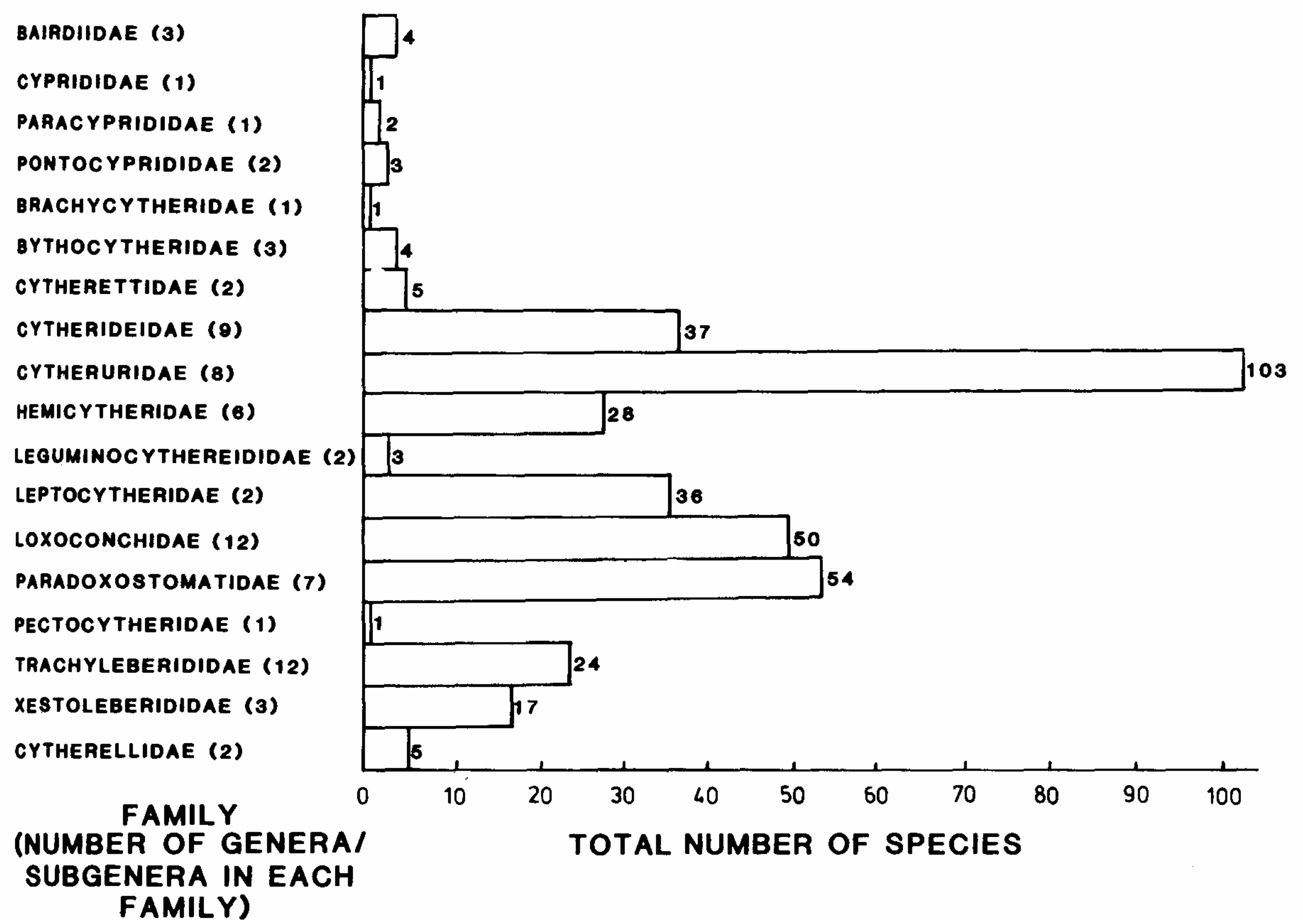

Figure 2. Species distribution by family: total number of families = 18, total number of genera/subgenera = 77, total number of species/subspecies = 378 and total number of specimens = 66,462.

The Ostracoda

The faunal composition of the St. Erth Ostracoda is shown in Figure' 2. The total number of species within each of the ostracod families occurring at St. Erth is indicated together with the number of genera in each family. The distribution of species within families is unusual when compared with other fossil and Recent, shallow water faunas. This is because the diversity of the Cytheruridae, Paradoxostomatidae, Loxonconchidae, Cytherideidae and Leptocytheridae is exceptionally high. The diversity of the Hemicytheridae is also high; but this family in other faunas often exhibits the highest diversity and it is, therefore, surprising to see it ranking sixth in order of species number at St. Erth. For the most part the distribution of the species within the various ostracod families is of palaeoecological significance. The 103 cytherurid species at St. Erth is indicative of the fine grained nature of the St. Erth lithology and low energy levels of deposition since this group contains most of the "small" ostracod species in the study. The high diversity of the Paradoxostomatidae with 54 species and to a lesser extent, the Xestoleberididae with 24 species, is indicative of a weed-rich palaeoenvironment as species of these families are commonly found on phytal substrates. Undoubtedly the high level of diversity attained by the Loxoconchidae, Cytherideidae, Leptocytheridae and Hemicytheridae is a reflection of niche availability or spatial heterogeneity in the comparatively shallow, warm water of the late Pliocene sea. The palaeoecological significance of the high number of leptocytherids at St. Erth has already been mentioned in connection with the possibility of there being lowered salinity régimes prevalent at St. Erth. The distribution of species in the families Trachyleberididae, Cytherettidae, Cytherellidae, Bairdiidae, Bythocytheridae, Pontocyprididae, Leguminocythereididae, Paracypridae, Cypridae, Brachycytheridae and Pectocytheridae is not remarkable as these are families which one would not expect to have achieved particular prominence in a late Pliocene, shallow water marine environment.

Comparison with Fossil and Recent Ostracod faunas

It is only when comparison with other faunas is made that the unique nature of the St. Erth deposits in terms of ostracod diversity, is appreciated. The Neogene to Recent time interval is chosen for comparative purposes as pre-Neogene ostracods differ at generic and sometimes familial level from those of the post-Palaeogene. The data is presented in tabular form (Table 1) and the comparative faunas have been selected primarily because they represent complete or very near complete faunas from which species and genera numbers could be readily ascertained. There are 3 obvious flaws concerning these comparisons: firstly that the majority of faunas concerned were not obtained from a single locality, like the St. Erth fauna. Secondly they do not represent equivalent time intervals and finally, in almost all cases, the number and weight of the samples from which the faunas were obtained are unknown. However, with respect to the inequality of the sample areas and time intervals compared, the diversity of the St. Erth fauna is indeed remarkable as only one sample locality is involved and the St. Erth deposits can be confidently assigned to a late stage in the Upper Pliocene (Jenkins, 1982 and Jenkins, Whittaker and Carlton, in press).

The most diverse of the comparative faunas indicated in Table 1 is the Redonian, Upper Pliocene fauna from North West France (Maybury and Whatley, in press a). This fauna was obtained from 58 samples and 16 localities. It comprises 384 species, 6 more than at St. Erth. There are certain parallels between the St. Erth. and the Redonian faunas, for example, individual samples yielded high numbers of species. A single 49 gram washed residue sample from Le Temple du Cerisier, one of the French localities, yielded 240 species. Also, there are 195 species (34%) of the total number of species from the two regions common to both. Given that the British and French faunas are approximately contemporaneous it is probable that the factors which influenced the diversity of the St. Erth fauna were also responsible for the diversity of the French fauna.

If the Redonian fauna of North West France is not taken into consideration, the most diverse of the fossil comparative faunas are from deep sea samples of the South West Pacific. Two-hundred and thirty-eight species from the Miocene, 200 from the Pliocene and 365 from the Quaternary have been recovered (Whatley, 1983). It must be emphasised, however, that these deep sea studies cover a latitudinal range of approximately 60° and involve the whole of the Miocene, Pliocene and Quaternary epochs, not just the upper part of the Pliocene as does the St. Erth study. Even when Recent faunas are considered, where taphonomic bias is not operative, diversity values do not approach those at St. Erth. In British coastal waters today, (down to 50m) there are about 120 species with the possible addition of 50 species if more open marine species are included (Whittaker lit. comm., 1985) and in tropical waters the diversity values are about the same with an average of 150-160 species recorded from any one area (Titterton, 1984 MS. and Titterton and Whatley, in press).

In addition, it is emphasised that the comparative faunas quoted in Table I are representative of the literature in that, with the exception of the French, Redonian fauna, no other ostracod fauna from any geographical interval was found either to equal or surpass the diversity of the St. Erth fauna.

Discussion

In order to account for the high.diversity of the St. Erth fauna it must first be considered whether this diversity is apparent or real. If the St. Erth beds represent a

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Age	Locality/Region	Author (date)	No. of Species	No. of Genera
Recent	Christchurh Harbour, The Fleet & Weymouth Bay, England	Whittaker (1972 MS.)	60	29
	Cardigan Bay, Southern Irish Sea	Wall (1969 MS.)	71	33
	Southern Irish Sea	Ralph (1983 MS.)	171	69
	Aracachon Bay & Bay of Biscay	Yassini (1969)	106	62
	The Rias of Pontevedra & Vigo, N.W. Spain	Ralph (1977 MS.)	61	39
	Mediterranean	Puri, Bonaduce & Gervasio (1969)	132	48
	Adriatic Sea	Bonaduce, Ciampo & Masoli (1975)	246	81
	Bou-Ismail Bay, West of Algiers	Yassini (1979)	115	54
	Guadalcanal & Shortland Island, Solomon Islands	Titterton (1984 MS.)	156	55
Ipswichian	Selsey, England	Whatley & Kaye (1971)	62	31
Quaternary	S.W. Pacific	Whatley (1983)	365	58
Late Pliocene	St. Erth, England	Maybury (1985 MS.)	378	77
	Kos, Greece	Mostafawi (1981)	89	42
Upper Pliocene	N.W. France	Maybury (1985 MS.)	384	86
	Coralline Crag, England	Wilkinson (1980)	62	42
Pliocene	S.E. France	Carbonnel & Ballesio (1982)	90	53
	S.W. Pacific	Whatley (1983)	200	50
Neogene (although predominantly Miocene)	Rhône Valley, France	Carbonnel (1969)	121	56
Miocene	Aquitaine Basin, France	Moyes (1965)	117	64.
	Fossil Beach, S.E. Australia	Whatley & Downing (1983)	98	44
	S.W. Pacific	Whatley (1983)	238	62

Table 1. The diversity of selected fossil and Recent Ostracod faunas.

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condensed sequence, enhanced diversity might be expected. There. is, however, no evidence for this as the ostracod and also the foraminiferan (see Margerel in Mitchell et al., op. cit.) faunas do not contain any contaminants. Both clearly indicate a late Pliocene age. In fact, on the basis of the presence of the planktonic foraminiferans, Globoratalia infiata and Neogloboquadrina pachyderma, Jenkins (1982) assigns a late Pliocene, G. infiata Zone age to the deposits.

Another possibility which would render the diversity of the St. Erth fauna apparent and not real is the possibility that the authors have failed to distinguish correctly between intraspecific and interspecific variation and, therefore, created more or fewer species than were actually present. The majority (88%) of the St. Erth Ostracoda are new. At present (1986) only 23 of these species have been formally described (Maybury and Whatley 1980, 1983a-b, 1984 and in press b and Whatley and Maybury 1983 and 1984). The entire fauna is, however, illustrated in Maybury (1985 MS.) and it is hoped to complete the publication of the fauna within the next 2 years, thus enabling the validity of the taxonomy to be scrutinised. We have shown the fauna to many colleagues. None would reduce our species list, some, however, suggest that our taxonomy is, in parts, over conservative!

In this study the cut-off point for picking the washed residue samples was the 100 mesh number sieve fraction (150/.tm) and it is noted that the most diverse genus that occurred at St. Erth is Semicytherura (with 58 species), which is very small. If the cut off point had been the 60 mesh number sieve fraction (215µm) as has been the case with several European and American studies, a valuable and quite considerable component of the fauna would be absent.

The St. Erth fauna contains many small, fragile specimens, particularly juveniles. Probably, therefore, taphonomy is a major factor contributing to the high diversity. The lack of derived material, careful sample preparation and accurate taxonomy combined with the favourable preservational history of the deposit have minimised the chances of the ostracod diversity being apparent.

Given that the abnormally high diversity is real and not apparent it is likely that a number of factors brought it about. The palaeoenvir0nmental conditions prevalent in the late Pliocene, as summarised above, were conducive to speciation. The sea was shallow and the palaeotemperature similar to that of the present day Mediterranean. The St. Erth fauna contains few extant species and most of these do not occur in British waters today, but in warmer waters. In the context of the Upper Pliocene, however, the St. Erth fauna appears to comprise a mixture of thermophilic and, to a lesser extent cryophilic species. The fact that both cooler and warmer water species seem to have co-existed in the same area is possible evidence for the proximity of an ameliorating, warm water current. The existence of a variety of niches for occupation, as evidenced by the presence of o stracods with a preference for phytal or other substrates and the differing salinity r6gimes obtaining during the deposition of the St. Erth clays, as indicated by the importance of the genus Leptocythere, are further factors which would promote diversification. Finally, if Catt and Wiers' hypothesis that the St. Erth sands where redeposited in transitory, coastal lagoons were true, the necessary requisite for allopatric speciation, a temporary barrier, would be present.

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Kaolinisation and isostatic readjustment in southwest England

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The fact that the china clay areas of south-west England occupy topographically high ground, yet have a poor resistance to erosion, and the interpretation of the origin of the china clay'that requires the passage of large volumes of water through the granite, while the processes actually cause a considerable decrease in permeability, are two problems which can be resolved by consideration of the isostatic movements that accompany kaolinisation..The alteration of granite to form china clay has involved a mass loss of about 20% at approximately.constant volume, causing a decrease in density from about 2.65 Mgm^{-3} for the fresh rock to about 2.23^{-3} on complete argillisation. Partial alteration gives rocks of intermediate density. A consequence of this decrease in density will be uplift to maintain isostatic equilibrium. For conservative figures of partial kaolinisation causing a reduction of density to $2.53Mgm^{-3}$ over a depth of lkm, the uplift is in the order of 50m. Within a 10km diameter pluton this uplift would give an overall dilation of 0.08m or widen joints by 80 µm every 10m. A cubic array of blocks of solid granite separated by joints with an aperture of 80 µn has a permeabilty of 8.5md. This is significantly greater than the limit of 0.5md needed for active hyclro-thermal circulation within abnormally radioactive plutons. Together with the observation that hydrothermal circulation is taking place within the granites of south-west England, it is suggested that this circulation indicates that kaolinisation is an active process, and that isostatic readjustment not only maintains the permeability of the granite as kaolinisation proceeds, but also helps explain the occurrence of the china clay in areas of high ground. The situation in south-west England can be contrasted with Brittany, where china clay areas occupy lower ground surrounded by the more resistant rocks of the metamorphic aureole, and kaolinisation is considered to be no longer an active process.

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Introduction

A curious feature of the topography of south-west England is the fact that the thoroughly kaolinised granite in the western part of the St. Austell pluton forms high ground. Indeed, the summit rises to a height of over 300m at Hensbarrow Downs between St. Dennis and Bugle. This is unusual because of the poor resistance to erosion possessed by the altered granite. Not even the presence of less easily eroded zones of relatively unaltered granite or the resistant quartz-tourmaline veins that cut the granite could account for this feature. The topography of the kaolinised granite appears even more unusual when it is compared with the ground at a height of about 200m occupied by fresh granite and the hornfels of the metamorphic aureole. Both fresh granite and hornfels are much more resistant to erosion than kaolinised granite. In the case of the kaolinised Variscan granites of Brittany, however, the topography is much as expected: with altered granite generally occupying low ground which is surrounded by hills formed of the fresh granite and metamorphic aureole. A good example in central Brittany is the Huelgoat granite.

Another difference between the kaolinised granites of south-west England and those of Brittany concerns their radiothermal characteristics. In Brittany the concentrations of radionuclides in the granites are comparatively small (4ppmU, 6ppm Th), so that these granites do not have a high heat generation capability - (Haenal, 1980). However, the granites of south-west England are abnormally radioactive (14ppmU, 15ppmTh) and are such a large source of radiogenic heat that hydrothermal convective circulation is taking place (Durrance et al., 1982; Burgess et al., 1982). Because kaolinisation is thought to be brought about during convective circulation of groundwater, it is possible that the topographic contrast between south-west England and Brittany is due to kaolinisation remaining an active process in south-west England, while in Brittany kaolinisation has ceased to operate.

Kaolinisation

The alteration of granite to china clay is generally considered to be merely the result of change of feldspar to

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kaolinite. But this alteration has not been a simple single-stage process nor, indeed, is it the only mineralogical 'change that has occurred within the granite. As a precursor to the formation of kaolinite, both Na and K feldspars in the south-west England granites were altered to smectite-illite and illite clay mineral assemblages, respectively, at a depth of about l-2km by high temperature waters (200-300°C) which were probably moderately to highly saline. This occurred some 260Ma ago. Only later, when low salinity waters penetrated the system did alteration of the remaining Na and K feldspars and smectite to kaolinite take place. The temperature of the water during this second stage of alteration was probably in the order of 50-150°C, and the process appears to have been operating at varying intensities over the last 200Ma, whenever the Cornubian peninsula was above sea level.

Accompanying the changes to the feldspars seen in the first stage of alteration, Fe in biotite became liberated and lost to the system. This is the stage of deferrugini-sation of the granite referred to by Durrance et al. (1982). Some mobilisation of Fe also occurred during the second stage of alteration, but in this process it appears to have been largely retained within the granite and gives rise to colour-staining of the rock. As colour-stained kaolinite is of little economic value, the best china clay deposits occur where the earlier stage of deferruginisation was most effective. Clearly, if the original mica in these areas was also deficient in Fe then the deferruginisation process need not have operated as effectively to still give good china clay deposits. Perhaps this is one reason why the main china clay deposits are associated with the Li-mica granites (such as in the western part of the St. Austell granite), but variations in the chemistry of the ground-waters producing the alteration may occur between biotite granite and Li-mica granite, so the true picture could be much more complex.

Finally, quartz within the granite also undergoes changes during both stages of alteration, but dissolution losses are somewhat off-set by the formation of SiO_2 overgrowths on quartz crystals elsewhere. In addition, some of the liberated SiO_2 gives rise to quartz veins. Although some SiO_2 liberated during argillisation thus remains within the system, some is also likely to have been removed.

As a result of these alterations, complete kaolinisation involves the loss of about 20% by weight of the granite, while volume change is comparatively small because of the presence of an interlocking framework of quartz crystals. Consequently, an unaltered granite with a density of about $2.65 Mgm^{-3}$ is reduced to a rock with a density of only about $2.2\ Mgm^{-3}$. However, where alteration has been only partially effective, then of course, intermediate density values result which reflect the degree of alteration suffered. Thoroughly kaolinised granite is known to extend to depths in excess of 250m, and, indeed, kaolinised zones within the granite have been noted at depths greater than lkm (A.V. Bromley, personal communication). Within the granite, the zones of alteration are clearly related to jointing. More extensively altered granite can either overlie or underlie less altered granite zones (stocks), and lateral changes from less altered to more altered granite are typical. Zone dimensions in the order of 10m or so are usually found and this is a typical value for the separation of joints below the zone of near-surface stress relief fractures (Heath, 1985).

Hydrothermal circulation

The removal of about 20% by weight of material from unaltered granite clearly requires the passage of a large volume of water through the system. Initially this flow must have been mainly confined to the most Open joints, although even in this case most of the water would probably have been concentrated in specific linear channelways (Heath and Durrance, 1985). Away from these joints, the rock mass as a whole eventually became altered by the ingress of water along grain boundaries. This implies that the intergranular permeability was significantly higher than the intact-rock permeabilities currently found in the granite below the zone of near-surface unloading (10^{-8}-10^{-9}d; Brace, 1980). Moreover, the alteration of feldspars to clay minerals is seen today as giving rise to reduced permeability in the surficial zone of alteraton of the granite. By analogy, and with the additional observation of the formation of quartz overgrowths, it is believed that both the first and second stages of alteration would have resulted in a circulation system which progressively became blocked unless energetic activation occurred to keep the flow routes open.

Energetic activation can be considered to comprise both forced movement of groundwater and the maintenance or opening of flow routes. Forced movement was considered by Durrance et al. (1982) to be hydrothermal convective circulation brought about by radiogenic heat in the granites in the manner proposed by Fehn et al. (1978) for the Conway granite of New Hampshire. Recently Fern (1985) has specifically modelled the granites of south-west England in terms of the radio-thermal convection of groundwater and demonstrated that there are no theoretical objections to such systems. Indeed, Gregory and Durrance (1985) have shown that the distributions of the radiogenic gases ^{4}He and ^{222}Rn in stream waters on the Carnmenellis granite closely match the observed heat flow. This is consistent with the rising limbs of groundwater convective circulation cells giving high radiogenic gas concentrations when they discharge into surface streams and localised high heat flow values. The circulation cells have dimensions in the order-of 2-3km and are thus of the form modelled by Fehn (1985).

The maintenance of an adequate permeability to permit extensive kaolinisation of the granite was considered by

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Durrance et al. (1982) to be largely dependent upon fracture reactivation at times of increased seismic activity. These were thought to be related to Mesozoic and Tertiary orogenic episodes. However, an analysis by Durrance (1985) recently has shown that isostatic movements can produce sufficient uplift to permit the opening of fractures to a very significant degree. In this context, kaolinisation processes which give rise to large mass' redistributions will, clearly, be accompanied by isostatic movements. It is the contribution these movements make to the maintenance of high permeability in the granite that is now investigated.

Isostatic movement and fracture reactivation

Before it is possible to determine the vertical uplift that will take place in order to re-establish isostatic equilibrium following the transfer of mass from a granite pluton on kaolinisation, we must first consider the two different models of isostasy proposed by Airy (1855) and Pratt (1858) to explain the differences between the astronomical and geodetic latitudes of survey stations in northern India observed by Pratt (1855). In the model of Pratt (1858), isostatic equilibrium is achieved when the heights of rock columns of different densities above a compensation level of uniform higher density are such that the products of the height and density for each column are constant. The model of Airy (1855), too, has a compensation level at the top of a layer of rock of uniformly high density, but the columns of rock above this level are made up of different lengths of lower and higher density material. In this case equilibrium conditions exist when the sums of the products of height and density for each column are constant (Heiskanen and Meinesz, 1958). In fact neither of these models is totally correct. Heiskanen and Meinesz (1958) consider that the lateral density variations of the Pratt model account for about 37% of the isostatic effect seen in a single area, while the vertical density variations of the Airy model are more important and account for the remaining 63% of the effect. Nevertheless, in south-west England the Pratt model is more appropriate: the low density granite batholith produces a local gravity anomaly with an amplitude of about -500gu and gives rise to the main topographic expression of the Cornubian peninsula and the high ground of the granite moorlands.

Redistribution of mass in either of these models produces isostatic disequilibrium. Consequently a displacement of the density-height relationships occurs in order to establish equilibrium once again. On the large scale such displacement probably takes place by flexure of the lithosphere and flow within the mantle below the lithosphere, especially where the lithosphere is a coherent unit. In south-west England, however, the response of the lithosphere to local change in load has been argued by Durrance (1985) to be governed by the presence of major fractures. This approach was based upon observation of severe local disruptions accompanying isostatic recovery on the retreat of the Devensian ice sheet in Scotland (Smith and Dawson, 1983). Local variations in uplift can be brought about by release of strain energy (Price, 1959) and recrystallisation at depth following removal of a superincumbent load.

In order to assess the overall change in density that is produced by kaolinisation, it must be assumed that only partial alteration of the granite is achieved down to a particular depth. For south-west England, conservative figures are for kaolinisation to reduce the density to $2.5Mgm^{-3}$ to a depth of lkm. For a cylindrical zone with a diameter of 10km this will involve the loss of about 10^{10} tonnes of material and is equivalent to the removal of an ice sheet 150m thick. The local gravitational anomaly produced by this cylinder is in the order of 50gu or approximately 10% of the gravity anomaly produced by the batholith as a whole.

In these circumstances, the uplift will be about 50m. This order of magnitude assessment is simply based upon the vertical uplift that is needed to restore isostatic equilibrium, and is obtained using the Pratt model together with the assumption that displacement occurs on vertical fractures within and around the altered zone.

If a slightly less conservative density decrease to $2.4Mgm^{-3}$ is considered to extend to a depth of lkm, the uplift is in the order of 80m. In fact, these figures probably more accurately represent the situation found in the St. Austell pluton. Indeed, a differential uplift of about 80m is consistent with the observed height difference of about 100m between the zone of kaolinised granite and the fresh granite of the St. Austell pluton. However, bearing in mind the need to be cautious as the constraining effect on uplift exercised by lithospheric flexure is uncertain, an uplift of 5om will be taken for subsequent analysis.

Because of the approximately spherical form of the Earth, any vertical uplift will be accompanied by a lateral extension (e), such that:

$e = dr/R$,

where d is the length of arc uplifted, r is the amount of uplift and R is the radius of the Earth. Therefore if we consider a pluton with a diameter of 10km having an uplift of 50m:

$e = 10^4 x 50/6.35x10^6 = 0.08m$

An extension of 0.08m in 10km is 80 μm in 10m. In other words, the uplift would open up by 80 μm those joints separated by 10m intervals of intact rocks.

Permeability

If we consider a cubic array of blocks of intact rock of side length d_1, separated by parallel-walled joints with an aperture d_2 which take laminar flow of groundwater, then where the intact rock permeability is negligible, the whole rock permeability (k) is given by Elder (1981) as:

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$k = (d_2)3/6d_l$.

If d_1 and d_2 are measured in metres then the units of k are m^2, Now $1d = 10^{-12}m^2$, so that:

$k(darcy) = (d_2)3x10^{12}/6d_l$,

where d_1 and d_2 are in metres.

Taking $d_1 = 10m$ and $d_2 = 80\mu m$:

$K = (80x10^{-6})\ x10^{12}/60 = 8.5md$.

A permeability of 8.5md may be compared with the value of 0.5md quoted by Fehn et al. (1978) as necessary for high flow rates of convective circulation to be established in abnormally radiothermal granites. For the less conservative figure of a density decrease to 2.4 M^{-3} to a depth of lkm and an uplift of 80m in a pluton of the same size, the resulting dilation would be 130µm every 10m and a permeability of 37md. Moreover, even if the calculations given above are repeated taking only the top 500m of granite to be partially kaolinised to an average density of $2.5Mgm^{-3}$, an uplift of about 25m would result, giving an. extension in a 10km diameter pluton of about 0.04m and a permeability for a 10m joint separation of lmd. Therefore, we may conclude that the isostatic uplift which accompanies kaolinisation is important in maintaining high permeability in the granite and the effectiveness of the kaolinisation processes.

Discussion

Although the main effect of the lateral extension that accompanies isostatic uplift would be the opening of preexisting joints, other phenomena may accompany this release of strain energy (Price, 1959). Now while small-scale movements of the kind considered above are unlikely to lead to the initiation of new fractures in granite (Durrance, 1985), Emery (1964) has noted that as rocks are polycrystalline granular materials the strain energy that is contained on grain boundaries as a result of tectonic deformation or crystallisation at depth in the Earth, is not released uniformly on unloading. Because different minerals possess different mechanical properties and different crystals can have a variety of orientations with respect to the load geometry, only some of the strain energy is lost instantaneously on uplift. The remainder of the strain energy is progressively released in response to changing environmental conditions, such as cyclical loading and unloading of the type used in the experiments of Price (1966). Only when the rock has become totally disaggregated has all the strain energy been released (Durrance, 1969).

Apart from cyclical loading and unloading stresses, changes of temperature, pressure, or chemical environment leading to solution/dissolution of minerals, can be responsible for inducing the release of strain energy. Clearly, hydrothermal circulation of groundwater through jointed rock will cause such changes to occur in the zones of intact rock between the joints. Thus it is apparent that once hydrothermal convective circulation begins to take place by fracture flow, then the intergranular permeability of the intact rock will be progressively increased as the hydrothermal system interacts with rock mass. This process will be most effective where the groundwater flow takes place along numerous small joints rather than in the occasional large aperture joints which may have spacings in the order of several hundred metres (Heath and Durrance, 1985). Because of the nature of the hydrostatic head produced by the thermal energy in convective circulation systems in fractured rock masses, the rising limb of a convection cell is characteristically related to a high flow rate pathway (Durrance, 1986). Conversely, drawdown within the circulation system is through the more widely distributed, closely-spaced joints. It is thus those rock masses affected by the drawdown sections of hydrothermal circulation systems which will undergo most release of intergranular strain energy and become more thoroughly altered. This is in agreement with the interpretation that envisages the second stage process of kaolinisation within the granites as occurring at sites of hydrothermal convective drawdown (Durrance et al., 1982), but also applies to the first stage of argillisation as well.

Conclusions

Alteration of granite leading to the formation of china clay deposits in south-west England has occurred in a two-stage process, involving groundwaters of different temperature and salinity. In both stages, alteration was brought about by hydrothermal convective circulation systems produced by radiogenic heat. Also in both stages the zones of alteration were located in the areas of convective drawdown. Differences in groundwater chemistry and temperature thus probably reflect the change in groundwater-envelope rock interaction and the depth at which the processes operated during the unroofing history of the granites. The maintenance of convection cells in the same area over a longperiod is not unexpected if groundwater circulation becomes locked onto major structural features such as faults. It is not surprising, therefore, to note the association of the main china clay deposits with the northwest-southeast transcurrent faults of south-west England.

Although argillisation of granite apparently produces a decrease in permeability, the mass loss that takes place at more-or-less constant volume results in a signficant isostatic uplift which increases joint apertures and increases the permeability of the system. Moreover, release of intergranular strain energy within unjointed zones of granite will be concentrated in areas of hydro-thermal drawdown and aid the alteration of. the granite. The argillisation processes themselves thus act to sustain hydrothermal circulation in the granites and the continuation of alteration. That this takes place today is supported by the observation of about 10ppm U in china clay from south Dartmoor, which is absorbed on the kaolinite. The china clay is, however, characterised by very low radiation levels, indicating that the U has not

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had time to develop an equilibrium series with its main gamma-ray emitting daughters (N.L. Jefferies, personal communication): the 238U decay series takes over 1Ma to achieve 99% equilibrium,

In so far as the second stage process of alteration is still taking place in south-west England, isostatic uplift could be sufficient to overcome the poor resistance to erosion possessed by thoroughly kaolinised granite and account for the topographic anomaly seen in the St. Austell pluton. The absence of radiothermal characteristics in the kaolinised granites of Brittany, however, suggests that other sources of heat were responsible if hydro-thermal circulation produced the china clay deposits of Huelgoat (Chauris, 1984). That no evidence of high heat flow is present in these areas of Brittany implies that kaolinisation is no longer an active process and that the topographic expression of the kaolinised granite and metamorphosed host rocks simply reflects their relative resistance to erosion. Indeed, the general character and limited thickness (10-20m) of the china clay deposits at Huelgoat are more consistent with an origin caused by deep weathering during the Tertiary (Bristow, 1977).

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Read at the Annual Conference of the Ussher Society, January 1986

Interaction between co-existing magmas: field evidence from the Cobo Granite and the Bordeaux Diorite Complex, north-west Guernsey, Channel Islands

R.S. D'LEMOS

D'Lemos, R.S. 1986. Interaction between co-existing magmas: field evidence from the Cobo Granite and the Bordeaux Diorite Complex, north-west Guernsey, Channel Islands. *Proceedings of the Ussher Society,* 6, 323-329.

The Cobo Granite of north-west Guernsey intrudes quartz diorite gneiss on its southern side and a diorite complex to the north and east. The phenomena observed at each contact differ markedly. At the granite/gneiss contact the granite clearly cuts the foliation of the gneiss, contains angular blocks of gneiss, has parallel-sided aplite offshoots and has produced only limited chemical changes in the gneiss. In contrast, a transitional zone occurs between the granite and diorite complex for which the term Marginal Facies is proposed. Members of the diorite complex in proximity to this zone contain unusually high abundances of K-feldspar, quartz and biotite. Enclaves of diorite within the Cobo Granite are typically rounded, though crenulate, lobate and pillowed types occur. Previous workers considered granitisation of solid diorite to have been the dominant process in this contact zone. This work indicates that a complex interaction between two magmas (one granitic, one dioritic)is a more plausible process to explain the phenomena observed.

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Introduction

Guernsey is the second largest of the Channel Islands, lying approximately 50km west of the French mainland of Normandy. The southern part of the island consists of a metamorphic complex (Fig. 1) comprising various para- and orthogneisses, thin metasedimentary screens and foliated igneous rocks, cut by minor intrusions of intermediate to basic composition. The age of the oldest rocks in the complex may be as much as 2000 Ma (Calvez and Vidal 1978), and these have been correlated with the Pentevrian basement of the adjacent French mainland (Roach 1977). To the north the metamorphic complex is intruded by an igneous complex comprising four major units, probably emplaced in the order St. Peter Port Gabbro, Bordeaux Diorite Complex, Cobo Granite, L'Ancresse Granodiorite, all of which are thought to be late Cadomian in age (Roach 1977).

The field relationships between two of these igneous units, the Cobo Granite and the Bordeaux Diorite Complex, are the mare subject of this paper. The relationships are far fromstraight-forward in two respects. First, there exists between the Cobo Granite and the Diorite Complex a zone of hybrid rocks for which the term Granite-Diorite Marginal Facies is proposed. Second, members of the Diorite complex in proximity to this zone contain unusually abundant alkali feldspar, quartz and biotite at the expense of amphibole and plagioclase. These features are interpreted as resulting from the interaction between the two igneous units. Drysdall (1957) interpreted such features as resulting from the intense granitisation, or metasomatism in the solid state, of

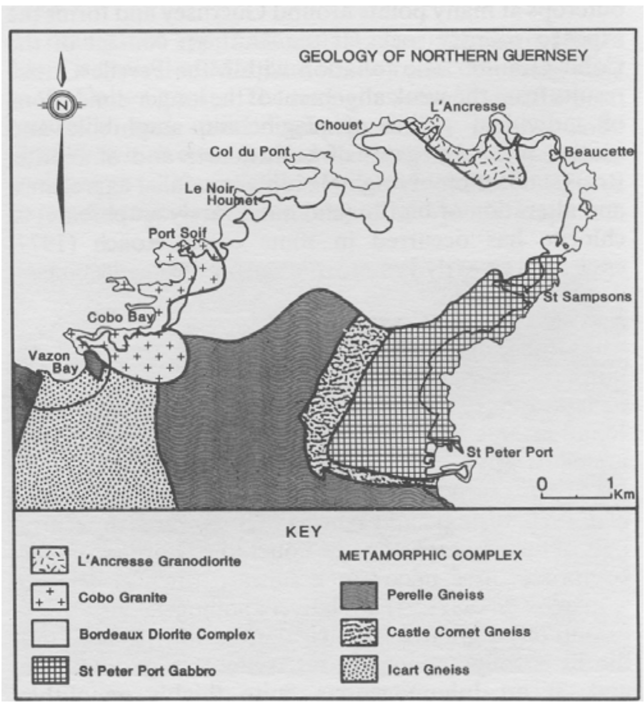

Figure 1. Geological map of Northern Guernsey (adapted after Roach 1966).

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Diorite Complex rocks, brought about by the intrusion of the Cobo Granite. This paper proposes a model involving two magmas, one granitic and one dioritic. The model involves complex magma-magma interaction and diffusion of mobile elements from granite into incompletely crystallised diorite magma. In order to illustrate some of the evidence for this model, field relationships between the Cobo Granite and Bordeaux Diorite Complex, and those between the Cobo Granite and the older Perelle Gneiss, will be described as well as a brief description of the major rock units involved.

The rock units

Cobo Granite

The Cobo Granite is a mineralogically, texturally and geochemically homogeneous granite (sensu stricto) which crops out almost continuously for 2.5km along the north-west coast of Guernsey. The granite consists of approximately equal proportions of partially altered plagioclase (1-2mm), quartz (2mm), and interstitial to poikilitic, flesh-coloured alkali feldspar (5-Smm), with a few percent of altered biotite and magnetite. Late stage aplites and pegmatites are absent from the granite, although restricted alkali feldspar rich zones associated with cataclastic bands do occur. Xenoliths occur sparsely throughout the body of the Cobo Granite, but are far more abundant towards the granite margins.

Perelle Gneiss

The Perelle Gneiss (Roach 1966) is a poorly to moderately foliated, medium-grained rock which varies in composition from quartz diorite to tonalite. The rock outcrops at many points around Guernsey and forms the exposed country rocks at the southern contact of the Cobo Granite. The foliation within the Perelle Gneiss results from the weak alignment of the longer dimensions of individual grains of plagioclase, amphibole and biotite, and aggregates of both quartz and of biotite. Recrystallisation of original biotite to smaller aggregates, and alteration of biotite (and more rarely amphibole) to chlorite has occurred in some cases. Roach (1977) envisaged an early Proterozoic age for the Perelle Gneiss.

Bordeaux Diorite Complex

The Bordeaux Diorite Complex comprises a large number of rock types ranging in composition from meladiorite to leucocratic tonalite and granodiorite. Relationships between rock types are complicated and are illustrative of a variety of petrogenetic processes. Layering is seen to occur in diorites at Beaucette (Elwell *et al* 1960, Bishop and French 1982). Towards the central part of the outcrop around Chouet, the Diorite Complex comprises three major rock suites (Brown *et al* 1980, Topley *et al* 1982): 1) a relatively homogeneous Diorite Group (even grained and acicular diorites), 2) a Grano-diorite Group (homogeneous, xenolithic granodiorite) and 3) an Inhomogenous Suite (highly xenolithic, Inhomogenous meladiorite to tonalite and granodiorite). Brown *et al* (1980) present geochemical data supporting this three-fold division, whilst Topley *et al* (1982) provide field evidence to show that although emplaced in the order I to 3 above, the Inhomogeneous Suite (3) was intruded at a time when the Diorite Group (1) was incompletely crystallised. The tripartite system proposed by Brown *et al* (1980) can be applied for the most part to those members of the Diorite Complex relevant to this work, though some minor differences do occur. The main differences are the occurrences of some slightly more mafic members in the Diorite Group, a sparsity of the Granodiorite Group, and occurrence of only the leucocratic members of the Inhomogeneous Suite. The most abundant members of the Diorite Group in this area are mesocratic diorites and quartz diorites. It is noteworthy that many of these rocks are similar in both mineralogy and geochemistry to the Perelle Gneiss.

Contact relationships

Cobo Granite/Perelle Gneiss

The contact between the Cobo Granite and the Perelle Gneiss is exposed in a number of reefs in the northern part of Vazon Bay (Fig. 1). The contact is somewhat sinuous and in places has been the site of later movement. Where exposed, the contact is sharp and the granite clearly cross-cuts the foliation and cataclastic bands within the Perelle Gneiss. At places along the contact the granite is sometimes finer-grained for a distance of several centimetres presumably as the result of chilling. Straight, parallel-sided aplite and porphyritic granite offshoots 50cm wide penetrate the Perelle Gneiss for a distance of up to 20m. In places along the contact alkali feldspar occurs within the Perelle Gneiss, although this is limited to a distance of centimetres to metres from the contact. Enclaves within the granite of the Perelle Gneiss and of associated dolerite dyke material are not generally abundant, but where they do occur they are typically angular and are dissected by brittle fractures into which granite has been injected. Megascopically such enclaves retain their original textures and do not appear to have undergone recrystallisation or growth of alkali feldspar porphyroblasts.

Taken as a whole, the field relationships at this contact are indicative of the intrusion of granite magma into completely crystalline, previously foliated country rock.

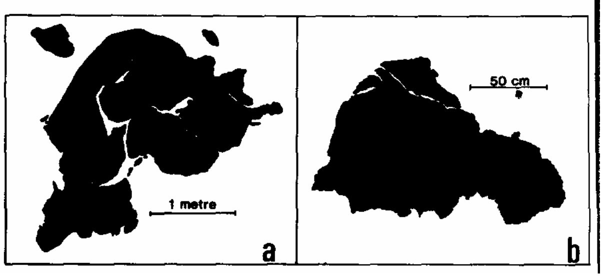

Figure 2. Lobate and pillowed enclaves with crenulate margins, west of Port Soif. (Traced from photograph, granite shown unornamented).

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Chemical interaction was sporadic in distribution and restricted to within a short distance of the contact, whilst mechafiical interaction resulted in the brittle fracture and local stoping of the country rock.

Coho Granite/Bordeaux Diorite Complex
As the margins of the Cobo Granite with the Diorite Complex are approached, both in Cobo Bay and Port Soft (Fig. 1), there is a dramatic and rapid increase in the size and, more particularly, abundance of xenoliths. This suggests that the xenoliths have been derived from the nearby Diorite Complex, although their present texture and mineralogy often differs from that of such diorites. The xenoliths, which may reach up to 3 metres in diameter, are typically drab green to black, fine grained (< 1.5mm) and consist of plagiodase, quartz, biotite and varying though generally small amounts of alkali feldspar. Xenoliths containing amphibole are rare, but where amphibole does occur it is an abundant constituent. The majority of xenoliths in the granite near the contact with the Diorite Complex have rounded forms with crenulate margins. In addition, highly lobate and pillowed enclave forms occur (Fig. 2) which have very fine-grained, crenulate, dark margins up to 10mm thick. One xenolith at Le Nic au Corbin, 200m west of Port Soif, measures 3m x 1.5m, has a rounded to pillowed form (Fig, 2a), and contains a vein which extends from the host granite into the interior of the xenolith. Within the xenolith, the vein branches and forms curving veins several centimetres thick, partially dividing the xenolith into a number of rounded, pillow-like forms. In addition to dark, fine-grained, crenulate margins along vein contacts, some contacts exhibit flame-like perturbations (Fig. 3) similar to those encountered in fluidised sediments.

Pillowed and lobate forms, crenulate margins, flame perturbations and curved irregular veins have been described from many localities throughout the world and have been interpreted as resulting from the interplay of two magmas (see for example Blake *et al* 1965, Walker and Skelhorn 1966, Wiebe 1973, 1980, Vogel 1982, Vogel *et al* 1984, Reid *et al* 1983, Key 1985, Bussell 1985). In many such cases involving two magmas, very fine-grained dark margins, and indeed the fine-grained nature of the enclaves as a whole, have been interpreted as resulting from the chilling and rapid crystal nucleation in hotter basic magma against cooler, less mafic granitic magma. However, in such intimate associations the interpretation of "dark margins" must be treated with caution (Bishop 1963, Topley *et al* 1982), and an alternative interpretation of the fine-grained nature of the enclaves could be that of recrystallisation.

The Granite-Diorite Marginal Facies
The most spectacular feature occurring at the contact between the Cobo Granite and the Diorite Complex is the presence of the Granite-Diorite Marginal Facies (hereafter referred to as Marginal Facies) which geographically separates the two major bodies. Two subfacies, or

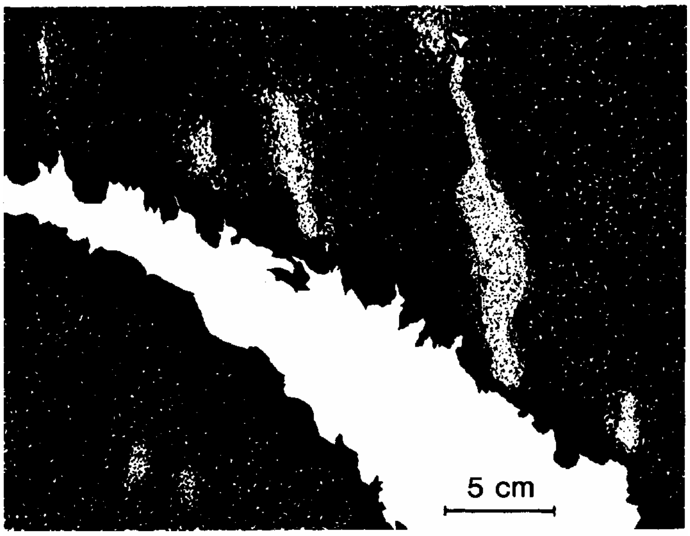

Figure 3. Flame-like structures and feldspathised zones associated with a small vein intruding pillowed enclave (Fig. 2a). Note the presence of a dark margin along the vein contact. (Traced from photograph, granite shown unornamented).

types of the Marginal Facies may be recognised; Type 1, a granitic rock containing abundant diffuse dark patches or clots up to 30mm in diameter: Type 2, a quartz monzonite containing skialiths (shadowy or ghostly xenolith). The two varieties have a number of characteristics in common in particular that the rocks are inhomogeneous, contain abundant enclaves which are always more basic than the host, and are intruded by granitic material.

Type 1
An example of Type 1 Marginal Facies is shown in Figure 4 and is typical of Marginal Facies rocks occurring close to the Cobo Granite. The rock consists of a granitic host, texturally and mineralogically similar to the Cobo Granite, which contains abundant (up to 50%) ill-defined enclaves up to 30mm in diameter to which the term clot (a purely descriptive term for a cluster of crystals) is applied.

Figure 4. Marginal Facies (Type 1). The granite host contains abundant diffuse dark clots (up to 30mm in diameter) and rare xenoliths.

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The clots are composed of fine plagioclase, biotite and minor quartz. The distribution of xenoliths and clots within the Marginal Facies is not random, but bears a close relationship to the proximity of the Cobo Granite. A gradational sequence from Cobo Granite into the Marginal Facies Type 1 may be identified in some areas of southeast Cobo Bay. This sequence may be summarised as follows:

1 Typical Cobo Granite containing rare xenoliths

2 Cobo Granite containing large, dark, fine-grained xenoliths. The xenoliths often contain sporadically distributed feldspar crystals which are larger than the grain size of the enclave as a whole.

3 Slightly darker, more mafic granite containing large enclaves, the interior portions of which are similar to the xenoliths outlined under (2) above, but which have poorly defined outer margins. The outer portion of such enclaves contains irregular diffuse patches of granitic material and darker diffuse patches, up to 30mm in diameter. These outer portions bear a marked similarity to the clot-rich marginal facies.

4 Irregular areas, several metres across, within granite which contain abundant dark clots, but within which fine-grained distinct xenolithic material is rare. At a few localities in southeast Cobo Bay, the clots have been drawn out into curved bands by local magmatic movement.

5 Large areas in which the clots are fairly evenly distributed and xenoliths are rare (Fig. 4).

In addition to the general gradation from granite into Marginal Facies outlined above, more abrupt contacts between Cobo Granite and Marginal Facies also occur, particularly where veins of Cobo Granite intrude into the Marginal Facies, though even these may be sharp or diffuse over a short distance. In some cases the contact is so straight and sharp that there can be little doubt that the Marginal Facies must have behaved in a brittle manner. In other instances veins are highly sinuous and irregular, have diffuse margins, and may grade into the clot-bearing rock along the vein length. Such relationships indicate that the matrix of the Marginal Facies was magmatic at the time of emplacement of the vein.

The larger feldspars occurring in the enclaves and the feldspar-rich outer portions of some enclaves may be interpreted as resulting from the metasomatic recrystallisation of the enclave (Drysdall 1957). A different interpretation is that some of the larger feldspars represent original phenocrysts in the material which now forms the enclave, or xenocrysts resulting from magma-mixing (see for example Vernon 1984, 1986). Interplay between the granite host and incompletely crystallised globules (Vernon 1984) of mafic magmas may have resulted in the diffuse outer portions observed in some enclaves. The clot-rich Marginal Facies is interpreted as resulting from the chemical and mechanical assimilation of large numbers of broadly dioritic enclaves (or possibly globules) into the Cobo Granite magma. The contacts between the Cobo Granite and the Marginal Facies show that the Granite was capable of intruding its own margins.

Type 2

The second type of Marginal Facies is gradational with Type 1. It is well exposed in Port Soif where it forms a zone some 20-50 metres wide separating Marginal Facies Type 1 from the Diorite Complex. In this area it contains few clots, but contains rounded skialiths and thin, irregular and diffuse granitic veinlets. The host material to the skialiths and veinlets is a variable mesocratic rock with a granite to quartz monzonite composition containing both biotite and amphibole. The skialiths are typically slightly darker, and less feldspathic than the host, whilst the veinlets are always lighter coloured. The presence of skialiths within the rock argues against a mode of formation involving the solid state metasomatism of a homogeneous diorite precursor. The relationships indicate the incorporation ofxenoliths into an inhomogeneous magma of quartz monzonitic composition.

At a greater distance from the Cobo Granite these rocks grade into homogeneous rocks of granodiorite to quartz monzodiorite composition containing 49% plagioclase, 15% quartz, 9% alkali feldspar, 13% amphibole and 13% biotite. The textures of such rocks in thin-section indicate crystallisation of all phases from a magma. Quartz and alkali feldspar are interstitial and may poikilitically enclose euhedral amphibole, zoned and partially corroded plagioclase, and biotite. Sphene, apatite and epidote are common accessory minerals. There is no evidence for growth of porphyroblasts (Vernon 1986), extensive recrystallisation or metasomatic replacement. The field relationships of these rocks, in view of investigations carried out between Port Soif and Le Noir Hoummet, and around Chouet (Topley *et al* 1982), strongly indicate that such rocks, despite the more granitic (sensu lato) composition, are genetically linked with the Diorite Group of the Bordeaux Diorite Complex into which they grade. A schematic section and summary of the field relationships at Port Soif is given in Fig. 5.

Discussion and Model

A model which requires metasomatic alteration of solid diorite as the dominant process in the formation of the features seen at the contact between the Cobo Granite and Bordeaux Diorite Complex is not acceptable on a number of grounds:

1. The contact relationships and interaction observed at the Cobo Granite/Diorite Complex contact differ markedly from those observed at the Cobo Granite/Perelle Gneiss contact. The similarity in geochemistry and mineralogy of the two country rocks (the Diorite Complex and the Perelle Gneiss), and the homogeneity of Cobo Granite, demonstrate that the processes occurring were not controlled by the composition of the country rock. Furthermore, the small size of the Cobo

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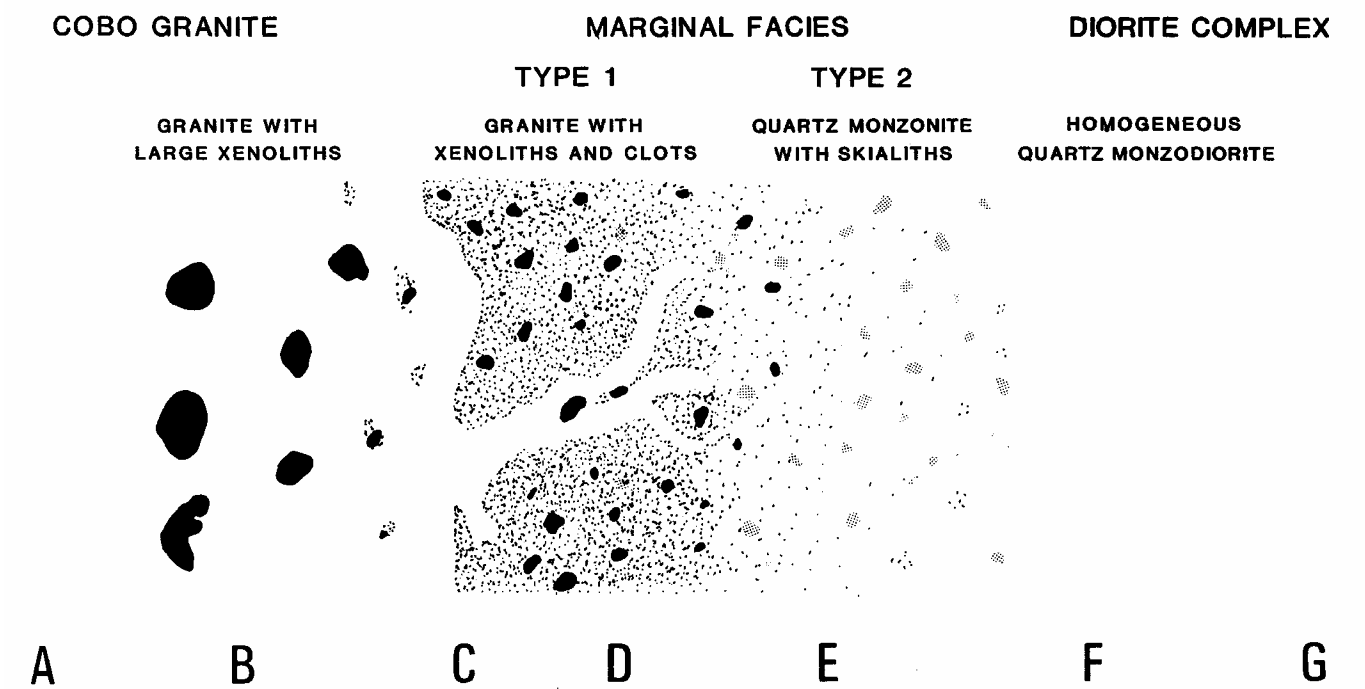

Figure 5. Schematic section across Port Soif. Approximate length of section=500m. A) Cobo Granite with sparse enclaves. B) Cobo Granite with abundant large enclaves, generally rounded but with occasional lobate and pillowed forms. C) A contact between Cobo Granite and Marginal Facies, gradational over a distance of a few centimetres to a metre. A more extensive gradational zone from Cobo Granite to Marginal Facies occurs in Cobo Bay. D) A xenolithic and clot-rich Marginal Facies formed by the break up and partial assimilation of large diorite enclaves. E) Inhomogeneous quartz monzonites which contain shadowy, dioritic skialiths and irregular and diffuse veinlets of granitic material. F) A gradation from E to homogeneous, apparently magmatic quartz monzodiorites which exhibit similar field relationships as diorites to the north. G) A gradation from F to unaffected diorites.

Granite requires that such differences cannot be the result of different depths, pressures or tectonic setting of the intrusion.

2. The extent of "alteration" of the diorites at the margins of the Cobo Granite (ie. the Marginal Facies) is considerably greater than the "alteration" of diorite xenoliths within the granite. If both the xenoliths and the marginal diorites were cool and solid, and had undergone metasomatism, it might be expected that both would show similar alteration.

3. The lobate and pillowed forms of some of the diorite enclaves within the Cobo Granite are not compatible with the incorporation of solid rock into a magma, but are more typical of the relationships seen between two magmas.

4. The wide area over which the "alteration" has occurred is probably not compatible with the process of metasomatism, given that metasomatism may be defined as mineralogical change resulting from the addition and/or removal of chemical components from a solid rock. Diffusion of chemical components is implicit in this process, but in crystalline silicates is known to be extremely slow (see Hofman 1980 for discussion)and over a geologically reasonable period might only account for metasomatic alteration over a number of centimetres to metres, rather than the zone of 100 metres, or more, seen in this case. Diffusion rates are greatly accelerated by the presence of obvious pathways such as shear zones, joint planes, cleavages, and brittle fractures, but in the present case there is no evidence for the existence of such pathways. In contast, diffusion rates in silicate melts are known to be far greater (Watson and Jurewicz 1984) than in solids. In addition, a small proportion of inter-crystalline melt will allow infiltration of magma in a similar fashion to that seen in some adcumulates. It is concluded that metasomatism, by diffusion of chemical components through solid, unfractured rock cannot account for the wide zone which has undergone enrichment in alkali feldspar and quartz. Combined magma-mixing, infiltration of magma and diffusion through a silicate melt is a more plausible process.

The Model

As a starting point, it is assumed that a homogeneous granite (Cobo Granite) was intruded into a previously foliated, completely crystalline quartz diorite gneiss (Perelle Gneiss). Brittle failure of the country rock accompanied limited stoping as is typical in such an environment. Chemical alteration occurred locally over only short distances. Towards the northern and eastern margins of the granite, the magma came into contact with a hotter, partially crystallised diorite magma which formed part of a large, broadly dioritic magma chamber (Bordeaux Diorite Complex). The initial process that

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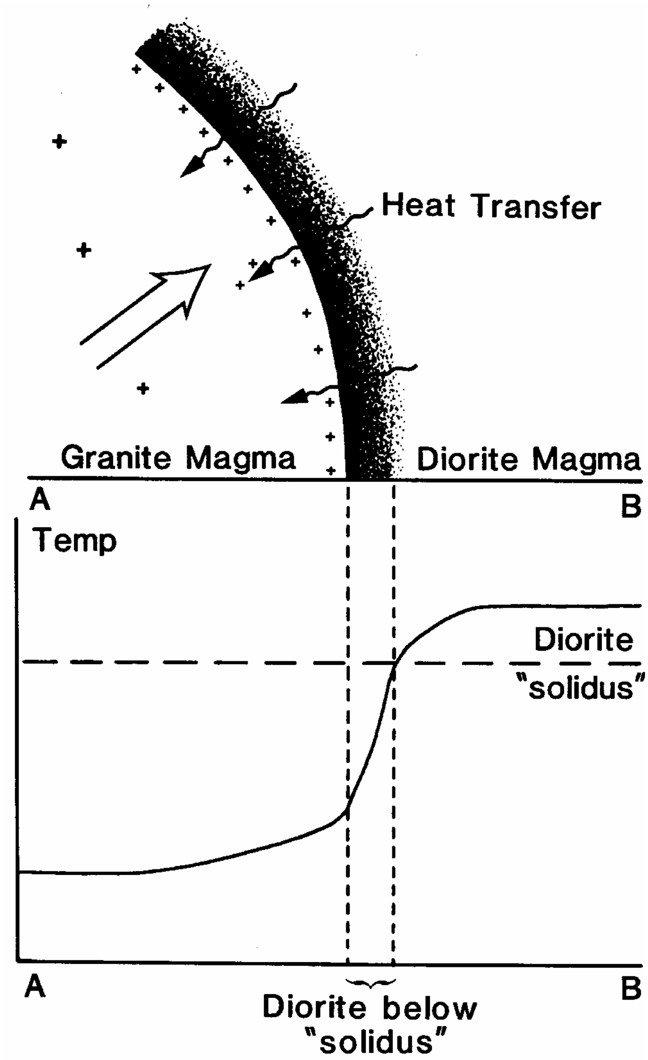

Figure 6. Formation of an essentially solid carapace by rapid chilling of hot diorite magma against cooler granite magma. The term "solidus" is used here to denote a temperature at which the diorite is sufficiently crystallised and viscous to resist flow and magmatic mixing.

occurred at this contact would have been the rapid transfer of heat from the hotter dioritic magma to the cooler, granite magma (Fig. 6). Undercooling of the dioritic magma would result in rapid nucleation of crystals and lead to the formation of a highly viscous, possibly solid, chilled carapace separating granite and diorite magma. Continued intrusion of the granite magma would tend to break up this carapace (Fig. 7), fragments of which would be incorporated as enclaves in the granite magma, and new carapace material, possibly of hybrid composition, might form. The process may have been continuous or may have occurred as a number of individual pulses. Lobate and pillowed enclaves might have been formed where more liquid portions of dioritic magma came into contact with granitic magma. It is proposed that where the proportion of granitic magma to dioritic magma was small, as at the extremities of vein systems penetrating through the carapace, the cooling

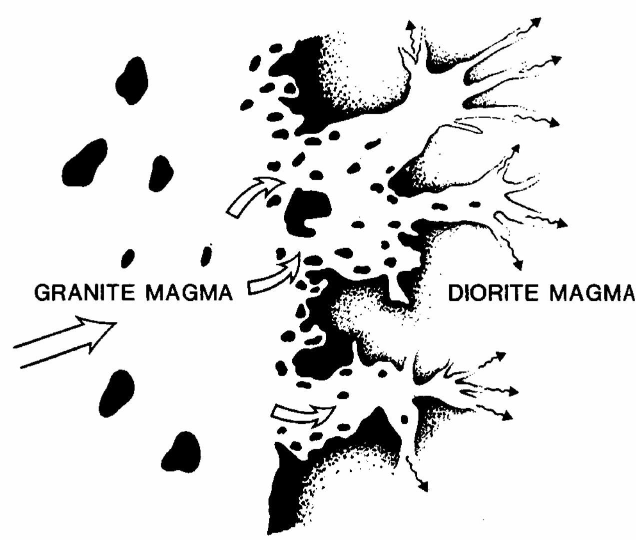

Figure 7. Break up and incorporation of carapace. Mixing occurs at the extremities of veins intruding the carapace, as denoted by small arrows.

effect of the granitic magma was negligible and magma mixing occurred. This gave rise to a magma compositionally between the two end members, ie. a quartz monzonite.

Over a period of time the temperature in the contact zone increased as heat was transferred from the dioritic body and thermal equilibration took place. Material produced by the break up of the carapace became distributed into both the granitic and mixed magmas. Enclaves within the cooler granitic magma were preserved, whilst those lying toward the hotter, dioritic side of the system underwent chemical and physical assimilation to form abundant clots. It is envisaged that over a period of time the system tended toward an equilibrium in that a smooth, gentle temperature gradient existed across the contact zone, and the earlier formed carapace became wholly broken up to form xenoliths towards the granite side and form partially assimilated xenoliths toward the dioritic side. Mixing between the two magmas was no longer restricted by a marked temperature contrast, but was inhibited by a wide zone of highly viscous enclave-rich Marginal Facies. This zone apparently formed a rheological barrier against continued mixing so that examples of true magma mixing are extremely rare. In this sense the Marginal Facies probably represents an early and arrested stage of the process of magma mixing. The only rock type which might possibly be viewed as the product of complete mixing is the quartz monzodiorite. However, the field relationships seen between this rock type and other members of the Bordeaux Dio. rite Complex, with which it is in contact, preclude magma mixing on a large scale. The author prefers the interpretation that the quartz monzodiorite formed by the infiltration and diffusion of such elements as silica and potassium into incompletely crystallised diorite magma during chemical equilibration across the contact zone.

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Conclusions

The field relationships described indicate that the Cobo Granite was emplaced into the incompletely crystallised Bordeaux Diorite Complex. The hotter, more mafic diorite magma chilled against the cooler granite magma to form a carapace which was subsequently broken up to form xenoliths within the granite and a xenolith and clot-rich Marginal Facies. True homogeneous magma mixing was very restricted due to the early formation of the carapace, and later, by a wide zone of partially crystallised Marginal Facies containing abundant solid material. Infiltration of small amounts of granitic melt and diffusion of mobile elements into incompletely crystallised diorite magma were probably the dominant processes involved in the formation of homogeneous quartz monzodiorite which is gradational between Marginal Facies and unaffected members of the Diorite Complex.

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Read at the Annual Conference of the Ussher Society, January 1986

Tin and tungsten in pelitic rocks from S.W. England and their behaviour in contact zones of granites and in mineralised areas.

K.E. BEER
T.K. BALL

Beer, K.E. and Ball, T.K. 1986. Tin and tungsten in pelitic rocks from S.W. England and their behaviour in contact zones of granites and in mineralised areas. *Proceedings of the Ussher Society, 6,* 330-337.

Background Sn and W levels in Cornubian pelites are 3-4ppm and 4-5ppm respectively. These elements show evidence for considerable primary dispersion around mineral deposits and all of the granitic cupolae studied. In granite aureoles the distribution patterns are consistent with the introduction by early metasomatic fluids and fixing in a front of Sn and W enrichment controlled by falling temperature. The zones of high values extend to at least a half kilometre from the contact, and indicate a notable crustal enrichment which predates the fissure filling mineralizations. Geochemical dispersion patterns surrounding vein mineralisation exhibit background populations dependent upon their positions within the aureoles.

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During the past decade the British Geological Survey has been conducting geochemical studies in various Cornubian mineralised areas, the results of which are now being published. At an early stage it became apparent that it was difficult to assign background values of metals to particular rock-types because of preferential sampling within mineralised zones. Furthermore, studies in the thermal aureoles surrounding the Sn- and W-enriched granites showed that the metamorphosed sediments were sometimes also enriched in those metals.

Several workers have reported Sn for SW England granites or porphyries but W determinations are sparse, though now improving. Floyd (1968) records Sn, but not W, from the metabasites and hornfelses of the Land's End aureole; the analytical method is presumed to be X-ray fluorescence spectrometry (XRFS). Until recently there were only two sets of published data for the Devonian sediments, Sn in slates and greywackes of the Perranporth area (Henley, 1974) using XRFS, and both Sn and W in "barren" killas from Carnmenellis (Hosking, 1964) using visual colorimetric methods. These results were later reconsidered by Edwards (1976).

This account gives details of Sn and W contents for various pelitic rocks from the area, and discusses the manner in which these values change with proximity to granite cusps and to mineralisation.

Geological setting.

A very simplified sketch map of the area is shown in Figure 1. Granitic rocks were intruded into folded Devonian and Carboniferous geosynclinal rocks which include a high proportion of mudstones. Only Devonian rocks are considered, which in the western part of the area are thought to represent Flysch type sedimentation, farther east the Devonian rocks are deeper-water mudstones, with included volcanic rocks and thick limestone lenses.

Traditionally three trends have been recognised in the metalliferous mineral veins in the area (Dines, 1956). The main trend is dominantly E-W or ENE-WSW with a southerly dip. This is frequently cut by so called "caunter lodes" of similar trend but opposing dip direction. Generally, but with important exceptions, these early lodes correspond to the regional strike. These lodes carry

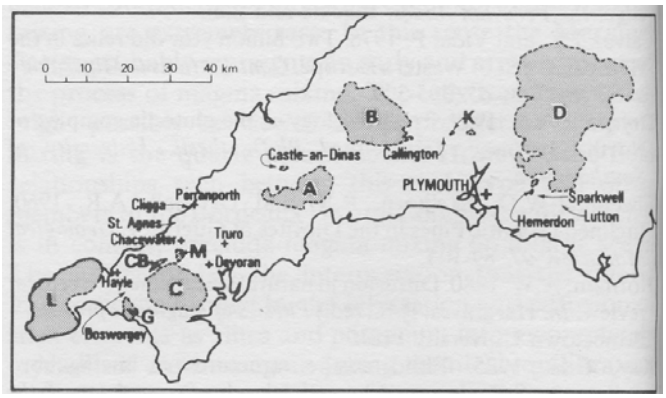

Figure 1. Location map for the areas studied.

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a high temperature mineral assemblage including important Sn and W mineralisation. A later mesothermal vein system' normally trending N or NW cuts, displaces and sometimes terminates the EW trending veins. Important relevant exceptions to these rules occur N of the St. Austell granite where a N-S- trending W vein is seen at Castle an Dinas. Sn and W mineralisation is most common in the early high temperature lodes and 500m either side of granite contacts. Economic W mineralisation is usually more tightly controlled within granites than the Sn mineralisation and important granite hosted stockworks are found in the Cligga Head and Hemerdon cusps.

Sampling locations

Pelites collected specifically as "background" samples were derived mainly from quarries and deep cuttings well removed from the granite margins in areas devoid of, or with only sparsely scattered mineral veining. Samples were collected from the Mylor Slate Formation near Hayle, Chacewater and Devoran, from the Gramscatho Group near St. Agnes and Truro, from slates associated with calc-silicate horizons in the Meadfoot Group near St. Columb Major, and from the Kate Brook and Brendon Formations near Callington. Thermal aureole samples were collected from surface exposures at Cligga and St. Agnes, from BGS boreholes at Bosworgey and commercial boreholes at Redmoor (Callington) and Hemerdon. Comparison is also made with pelites collected during BGS geochemical investigations at Royalton, Castle-an-Dinas, Mulberry, Lutton and Sparkwell (Figs. 1 and 7).

Analytical procedures.

Sn and W were mostly determined by XRFS. Because of the requirement for a speedy throughput, and the high values frequent in mineralised areas, a rapid XRFS procedure was employed with high detection limits of 5 and sometimes 10ppm. A more time-consuming but precise XRFS method (Philips PW 1450/10 AHP, Mo tube, LiF200, 30KV and 60mA, Smith, 1984) with detection limits (3xs.d.) of about lppm were used for the aureole and background rocks, supplemented in some cases by neutron activation analysis (NAA) for W. Unfortunately there are few international rock reference standards for Sn and W with values in the range of interest for the rocks of SW England. Some indication of accuracy can be obtained by comparison of XRFS and NAA for W where a product moment correlation coefficient (r) of 0.92 was obtained for 20 samples ranging 4-20ppm. Comparison with a colourimetric method for W in the range 120-650ppm gave a value for r of 0.885. Standards were also checked by spiking with wolframite and SnO_2.

For aureole and background samples and where samples from cored boreholes were available, the pelitic nature of the rocks was confirmed by petrographic examination and multielement geochemical analysis. In some mineralised areas examination of the rocks was largely confined to careful examination of washed chippings using binocular microscopes. There is no confirmatory major element data.

Table 1 Tin concentrations in pelites

LOCALITY	SOURCE	CONCENTRATION
MUDSTONES		
World	Onishi and Sandell, 1957	4ppm [67]
N. America	Hamaguchi and	11ppm
Japan	Kuroda, 1969	2.5ppm
Europe		5ppm
Japan	Terashima and Ishihara, 1982	2.9ppm [51]
Perranporth	Henley, 1974	17ppm (med.) [96]
Gramscatho		18ppm (med.)
Penhale and Ligger		13ppm (med.)
Carnmenellis	Hocking, 1964	10ppm (med.) [26]
MUDS		
Oceanic pelagic	Hamaguchi et al., 1964	0.8-7.5ppm [13]
	El Wakeel and Riley, 1969	10-20ppm [18]
	Barsukov et al., 1975	2ppm [12]
	Smith and Burton, 1972	3.4ppm [12]
SANDSTONES		
Gramscatho	Henley, 1974	17ppm (med.)
Cheri banded slates		22pm (med.)
PELITES	THIS WORK	3ppm (med.) [22] 3.3ppm (mean) 4.4ppm (s.d.)

med. =median. s.d. = standard deviation. [ ] = number of analyses, where given.

Determination of background levels.

Because there is clear petrographic evidence for thermal alteration extending at least 1km from the granite contacts, and inherent uncertainties in assessing the depth to buried granite contacts by geophysical methods, samples for this part of the study were confined to distances of at least 2km from mapped contacts.

Table 1 summarises the data obtained for Sn in the present survey and compares these values with those Cornish results used by Edwards (1976) and with published world means. Henley (1974) includes within his samples several from well within the Cligga Head Granite aureole, which our examination shows to be Sn-enriched (see below). Even allowing for such inflated values, however, it seems that Henley's mean background figure would be at least twice that of ours. There is a similar relationship with the results of Hosking (1964), this time

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Table 2. Tungsten concentrations in pelites

LOCALITY	SOURCE	CONCENTRATION
MUDSTONES		
World	Turekian and Wedepohl, 1961	1.8ppm [c]
	Vinogradov, 1962	2.0ppm [c]
	Horn and Adams, 1966	1.9ppm [c]
Russia	Vinogradov et al., 1958	1.8ppm [c]
India	Dekate, 1967	1.5ppm
Uganda	Jeffery, 1959	3.8ppm [23]
N. America	Harmon et al., 1978	3.2ppm [26]
MUDS		
Okhotsk Sea	Isayeva, 1960	5-70ppm
Kamchatka	Petelin and Ostroumow, 1961	15-40ppm
Black Sea	Pilipchuk and Volkov, 1966	10-13ppm
PELITES	THIS WORK	4ppm (med.) [22] 4.6ppm (mean) 4.3ppm (s.d.)

med. = median. s.d. = standard deviation. [ ] = number of analyses, where given. [c] = data based on compilation. Data from Vinogradov et al. is based upon a composite of 7,614 samples.

by a factor of 3. It may be suspected that his samples, some reporting 100ppm, are not strictly "barren" or "background" but are affected by the proximity of the Carnmenellis Granite. Furthermore, there is an added degree of uncertainty introduced by comparing instrumental results with those determined by visual colorimetry. Our mean, at 3.3ppm Sn, is also somewhat lower than the calculated world mean although individual values, < 1 to 26ppm, fall within the quoted ranges of most authors. Lister (1984) also records two values of Sn in pelites of < 3ppm.

Table 2 gives details of W in Cornubian mudstones and compares the values with determinations by various authors. From a range of 2 to 8ppm our mean calculates at 4.6ppm which is about twice the world average. The values are close to those obtained from mineralised areas in Uganda and overlaps values for pelagic muds from the Okhotsk Sea and Kamchatka area. Isayeva (1960) postulated that proximity to volcanoes might explain the high values of W in muds from the former basin. Harmon et al (1978) indicated an ill defined but nonetheless significant correlation between organic content and W in pelitic rocks from N. America. Lister (1984) recorded W values of 2ppm for two background slates. Hosking's "barren" slates were analysed colorimetrically for W giving a mean of 7.2ppm and median value of 4ppm from a set ranging from 2 to 30ppm. Again our results give a lower mean figure but all individual values fall within Hosking's range.

In summary, therefore, the Sn content of unmineralised pelites outside the likely zones of influence of the granites is rather less than world averages, although within their quoted ranges, whilst the W concentration is about twice the world average, this perhaps related to contemporaneous volcanic activity or the almost ubiquitous content of organic carbon.

Granite aureoles.

The five examples used in this study surround cuspate or ridge-like parts of the batholith where sampling is possible from outcrop - Cligga and St. Agnes - or from extensive drilling - Bosworgey, Redmoor (Callington) and Hemerdon (see Fig. 1). All of these granites are specialised for Sn and it has proved impossible to find a suitably low-Sn granite in Cornubia for comparative study. Low W-granites, however, have been recognised and in this section of the paper the investigations into the aureole geochemistry are described in order of increasing W.

Drilling at Redmoor has proved under Kelly Bray village the concealed westerly extension of the Hingston Down - Kit Hill granite ridge, the apex of which is here indicated as plunging gently westwards. Aureole samples were taken southwards from this ridge, mainly from company boreholes. The mean Sn value of 10 granite specimens is 27ppm, while that for W is 7ppm. Though W is low by SW England standards, the concentration is still high on a world basis (e.g. Tischendorf, 1974). The aureole rocks exhibit similar patterns of both Sn and W distribution (Fig. 2). High and variable values of Sn extend between 50 and 500m from the contact, but at 600m the values decrease and at about 1km are close to the background levels at less than 10ppm. For W the metal values are much lower but a zone of high values extends to about

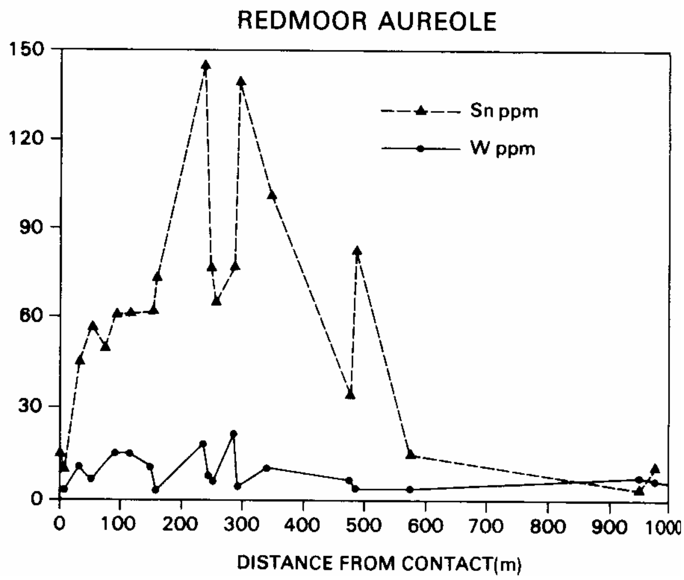

Figure 2. Geochemical profile for Sn (dashed lines and triangles) and W (full lines and circles) for the Redmoor aureole.

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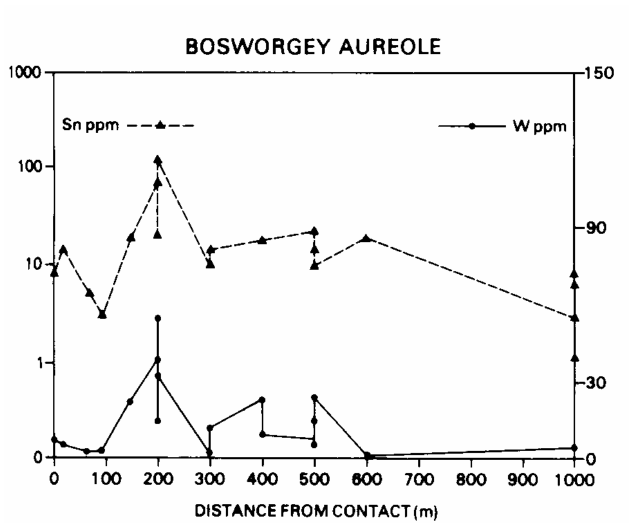

Figure 3. Geochemical profile for Sn (dashed lines and triangles) and W (full lines and circles) for the Bosworgey granite. Note that the Sn axis is logarithmic.

400m before declining to near background levels farther out.

The Bosworgey Granite is a concealed cuspate body located near to one of the emanative centres of Dines (1934) at the intersection of the projected Tregonning - Godolphin and Carn Brea- Carn Arthen trends. The geo-chemistry of the granite was described by Ball and Basham (1984), who reported mean values of 71ppm Sn and 19ppm W. The aureole rocks were sampled by core and percussion drillholes. The Sn and W distributions within the slate envelope are similar to each other and to that observed in the Redmoor aureole. The Sn pattern (Fig. 3) shows high values extending from 100m to 500m from the contact with lower but still anomalous values near the contact and close to background levels farther out. The W levels show a similar distribution, the absolute concentrations being higher than those at Redmoor.

At St. Agnes the granite is only just unroofed and the best exposures are confined to Cameron Quarry, recently described by Hosking and Camm (1985). Values for Sn of more than 1000ppm are common and the mean W value for 7 samples is 25ppm. In the aureole high values of Sn extend to at least 50om, and maybe 80om, from the contact (Fig. 4) though the distribution is affected by some significant lithological inhomogeneity. Again, concentrations close to the granite are lower. In close similarity to the other two aureoles, W is high to about 400m and declines to background levels at 1km from the granite.

The Cligga Head Granite differs from the others described in that the cusp is heavily mineralised by sheeted greisen veins carrying wolframite and less abundant cassiterite. The geochemistry is described by Hall (1971). Overall Sn and W levels in the granite are, in

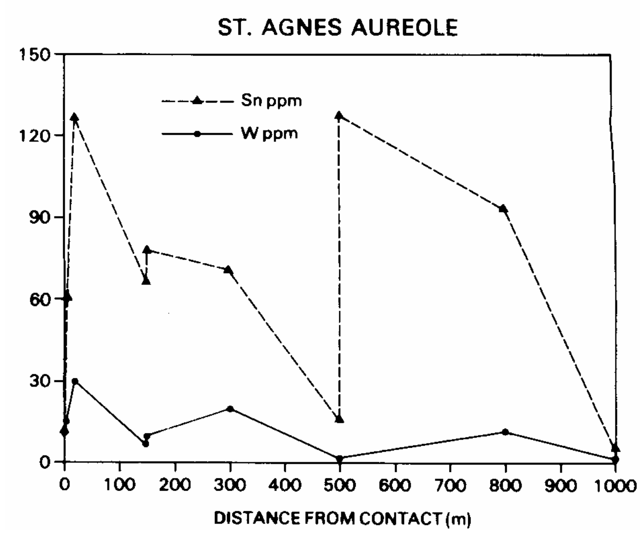

Figure 4. Geochemical profile for Sn (dashed lines and triangles) and W (full lines and circles) for the St. Agnes aureole.

consequence, difficult to determine with accuracy but unmineralised samples typically have values of about 30ppm Sn and 100ppm W. The interpretation of the aureole geochemistry is also complicated by the rocks being heterogeneous; there are greater concentrations of chert banded slates in the outer part of the aureole. Sn values are high close to the granite (Fig. 5) and this zone of high values extends outwards for 600 or 800m. Similarly the W values are high close to the granite and tail off to background concentrations at 800m. A similar distribution was observed for Sn by Henley (1970).

The final example, the Hemerdon Granite, is also heavily mineralised. A stockwork of interconnected quartz, quartz- feldspar and greisen-bordered quartz veins

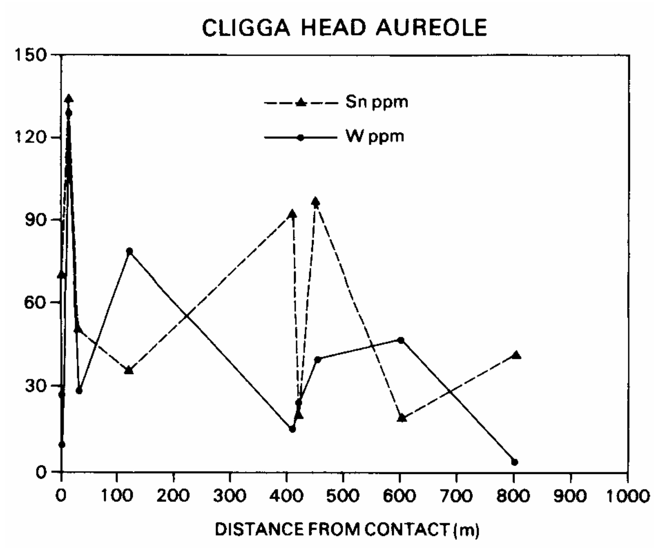

Figure 5. Geochemical profile for Sn (dashed lines and triangles) and W (full lines and circles) for the Cligga Head aureole.

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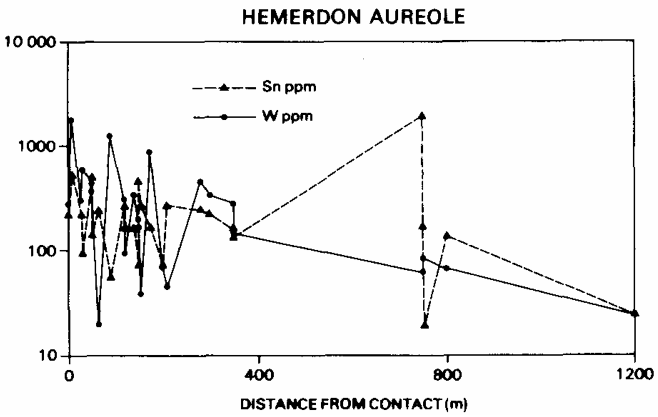

Figure 6. Geochemical profile for Sn (dashed lines and triangles) and W (full lines and circles) for the Hemerdon aureole.

bearing occasional large blades or clusters of wolframite is developed within the near-vertical granite dyke with some veins extending short distances into the inner part of the aureole. Unlike the foregoing four examples there is here no marked Sn or W relative depletion at the contact (Fig. 6) and even at 1200m these metals have not dropped to true background levels. This in part may be due to interference from the aureole of the Crown Hill Granite immediately to the north. However, it is pertinent to note that apparently unmineralised core from the Sparkwell (Great Stero boreholes, some 2km ESE of Hemerdon, gives a normal background of 5ppm for W but 70% of the Sn values were above background (Table 3). At the Lutton boreholes, about 2.5km from Hemerdon, there is a large population averaging 50ppm Sn. It seems then that the W levels must drop to background somewhere beyond 1200m from the contact; Sn remains high in the surrounding area and at Lutton is apparently related to the abundance of greenstone (Beer *et al,* 1981).

Despite these wider spread high levels of both metals there is in the Sn plot (Fig. 6) evidence of a marked change of general tenor at about 750m from the contact. In the case of W there is only a minor interruption in the downward trend at that distance.

The common pattern, then, is for a narrow (20-50m)exo-contact zone of moderate Sn and W values, succeeded outwards by variably high metal values to a distance of usually 500 to 800m, after which the values fall rapidly to background. This pattern we interpret as indicative of an early pervasive form of primary metal dispersion carried outward from the solidifying granite "hood" by metasomatic fluids and depositionally controlled by an interaction of rock and fluid chemistry in a magmatically imposed temperature gradient. Such a dispersion, frozen as a "front" around the granite, is totally unrelated to later mineral veining. Dispersion haloes developed around later mineralisation, be it metasomatic (greisen associated) or hydrothermal, may commonly be super-imposed on the primary front, modifying its apparent

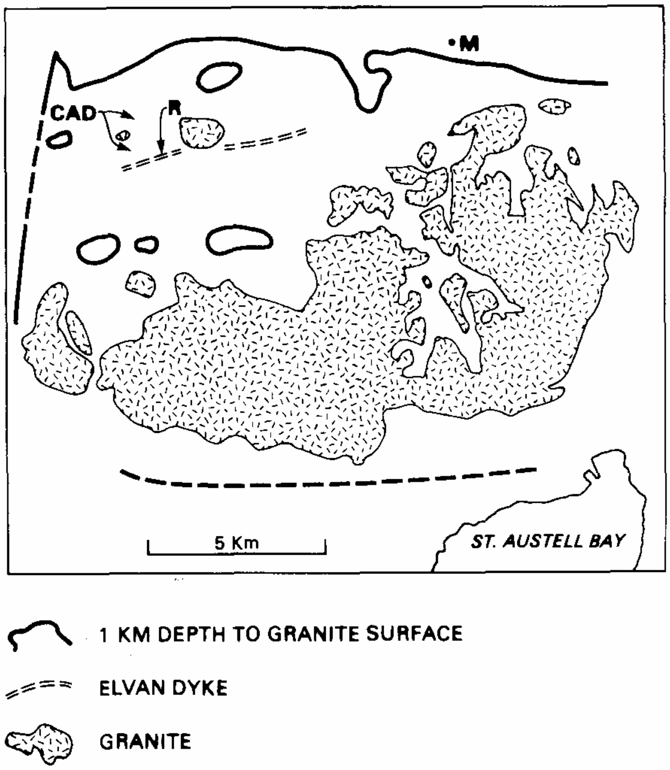

Figure 7. Locality map of the St. Austell area showing positions of the various granites. CAD = Castle an Dinas areas. R Royalton. M = Mulberry.

extent and its detailed tenor. It is believed that such a model explains the differing distribution evident at Hemerdon.

Mineralised areas.

Data from BGS studies of Sn- or W-mineralised areas were examined to ascertain whether the background levels of these metals were still recognisable in the pelitic host rocks.

A swarm of narrow tourmaline-cassiterite veinlets cutting a granite-porphyry dyke was formerly worked at Royalton (Dines, 1956), south-west of the Belowda Granite outcrop (Fig. 7). Rotary percussion drilling was undertaken in the slates to the north of the openwork and samples taken at fixed intervals down-hole were analysed by XRFS for Sn, but W was not determined (Beer, Turton and Ball, 1986).

Figure 8 shows a log-probability plot (see e.g. Sinclair, 1976) for 280 drill samples. Three log-normally distributed populations can be distinguished separated by inflexion points represented by crosses. Each population can be represented graphically by the straight lines in which the marks represent means + 1 s.d. The lowest population represents only 6% of the data and is affected by truncation at 5ppm, so that only a rough estimate of the mean of this population may be made. The main population (84%) has a mean of 1 lppm, while the highest

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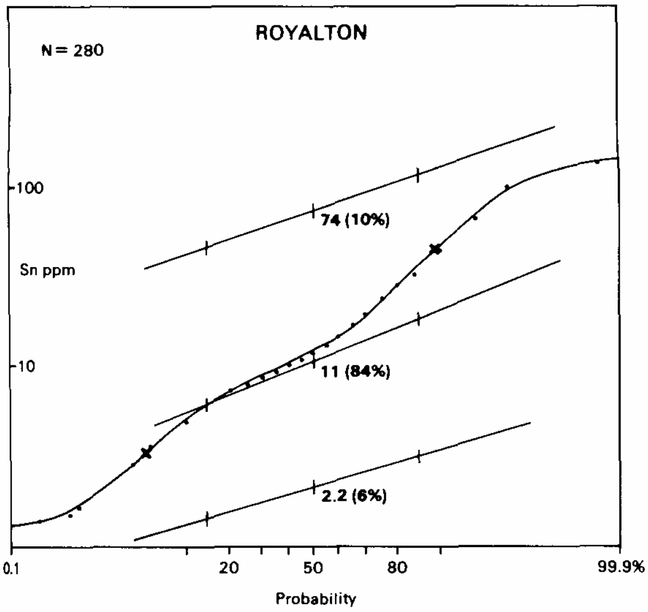

Figure 8. Log. probability plot for pelites from the Royalton area.

Table 3 Population statistics

Locality		No.	Lowest population	Other populations
Royalton	Sn	280	2.2(6%)	11(84%), 74(10%)
	W	no data		
Castle an Dinas S	Sn	509	<10(2%)	56(49%), 260(49%)
	W		2-5(60%)	23(40%)
Castle an Dinas N	Sn	487	10(50%)	100(50%)
	W	20(100%)		
Mulberry	Sn	15	102(100%)	
	W		5.5(75%)	43(25%)
Lutton	Sn	19	2.5(15%)	50(85%)
	W		56(100%)	
Sparkwell	Sn	9	5(30%)	22(70%)
	W		5(90%)	21(10%)

Population data in the form: mean (% of total population). Analysis based only on probability plots for the first three localities. Confirmatory histograms were produced for Mulberry and Lutton because of the small sample size. Histograms alone were used for the Sparkwell data.

group has a mean of 74ppm and represents some 10% of the data from this locality.

A short distance to the west similar drilling was carried out over an area assumed to contain southerly extensions of the Castle-an-Dinas Wolfram Lode (Beer, Ball and Bennett, 1986). Probability plot analysis provides the data summarised in Table 3. A regional background population for Sn (< 10ppm) is barely detectable whereas that for W, at 2-5ppm, comprises a major part of the total data.

The northern extension of the same lode was examined in a separate drilling programme. Only one population was recognised for W, with a mean of 20ppm; two Sn populations were indicated with means of 10 and 100ppm.

These three areas are close to the roof of the St. Austell Granite and well within the contour marking 1000m to the granite contact as calculated by Tombs (1977). It is not surprising, therefore, that high local backgrounds are observed, these also being consistent with observations from the distribution patterns of Sn and W in aureole rocks. Even so, there emerge from some areas vestigial indications of a lower, regional background.

A comparison has been made by examining data from areas at more than 1km from a granite contact. At Mulberry and nearby Wheal Prosper (Fig. 7) large sheeted vein swarms in slate were intermittently worked for cassiterite. A little wolframite is known from both locations but there is no recorded production of tungsten. Drilling carried out near the intersection of the two vein trends (Bennett and Beer, 1981) provided the data quoted in Table 3. A majority of the W results belong to the regional background population whereas all the Sn results fall into a single "high" set. Clearly, here the dispersion halo from the Sn mineralisation completely swamps the background levels.

At Lutton on the other hand, where the drillhole intersects a W-mineralised structure (Beer *et al.,* 1981), there is only one population for W. The regional background levels are statistically overwhelmed by the mineralisation dispersion, though low values within the set almost certainly reflect that background. Sn shows a clear bimodality, a small (15%) population being scattered around a 2.5ppm Sn mean background level.

Nine samples taken from boreholes at Great Stert, Sparkwell (Fig. 1) yield results which plot to show apparently bimodal distributions of both Sn and W. The low value populations accord well with the regional backgrounds previously derived. In the case of Sn this is heavily overprinted by a halo, the source for which is not obvious.

Discussion

The distribution of both Sn and W in relation to granite plutons is similar. Usually there is an inner zone of moderately high values extending to 50 or 100m from the contact and, outside this region, a zone extending to approximately 500m in which higher but irregular concentrations are observed. Beyond this intermediate zone there is a gradual decline in metal tenor and levels indistinguishable from background are recorded beyond 600 to 800m. These distances are only approximate and appear to vary from case to case. The size of the W maximum at intermediate distances appears to depend

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upon the concentration of W in the associated pluton, and a similar but less certain case may be made for Sn. Where the pluton itself is mineralised the innermost low value zone is usually obscured or overprinted. For the same concentration in the pluton, Sn is more widely dispersed than W. Henley (1970) for Sn, and Lister (1984) for Sn and W, also showed that there was an increase in these metal contents with approach to the contacts. The latter author indicated that both Sn and W reached maximum concentrations in xenoliths.

This distribution can be explained simply in terms of mobilisation and dispersion of the metals from the granite and the hot inner part of the aureole with eventual fixing in a lower temperature region. However, we favour a modified explanation which takes into account observed alteration in the aureole and which may explain at least some of the observed variations.

Typically the inner part of the aureole, i.e. up to about 50m or so from the contact, is characterised by the growth of feldspar: it is in turn succeeded by a biotite zone to about 200m and then by a zone of spotting to about 1km. The feldspar zone appears to resist the later metasomatic and hydrothermal alteration which is pervasive and common in both the biotite zone and the inner part of the spotted zone. This alteration consists variously of fluoritisation, .tourmalinisation and greisening, all of which facilitate or accompany the deposition of cassiterite and wolframite. It is tentatively suggested that many of the high values in the thermal aureoles can be attributed to such controls working on primary Sn and W dispersed into the envelope rocks during the early phases of granite consolidation.

Conclusions.

The background levels of Sn and W for pelitic rocks in SW England are unexceptional. The levels are lower than the world averages for Sn but approximately twice the world average for W.

Pelitic rocks in the thermal aureoles of the granite plutons all show consistent enrichments in these metals, the extent of the enrichments being crudely dependent upon the degree of specialisation of the associated granite. Some of this enrichment is attributable to early metasomatic alteration of a type similar to that which affects the granites and which frequently results in increased Sn and W concentrations in those granites. The zones of enhanced values extend to at least 0.5km around the plutons indicating a major introduction of both Sn and W at high crustal levels, apparently in advance of the main fissure filling mineralizations.

There is clear evidence for considerable primary dispersion of these metals around the vein deposits. Usually, background populations can still be identified within the geochemical data but the level of metals within the enriched "dispersion" populations may be related to their specific geological settings. For example, deposits set within the thermal aureole of a granite always have higher local background values than those outside the aureole.

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Read at the Annual Conference of the Ussher Society, January 1986

Mineralogy and illite crystallinity of the pelitic Devonian and Carboniferous strata of north Devon and western Somerset.

U. KELM

Kelm, U. 1986. Mineralogy and illite crystallinity of the pelitic Devonian and Carboniferous strata of north Devon and western Somerset. *Proceedings of the Ussher Society,* 6; 338-343.

A preliminary map of the metamorphic zones in the pre-Permian strata in North Devon and West Somerset is presented. The metamorphism varies from mid-diagenesis to epizone and bo measure ments indicate a low pressure regime. In the diagenetic field variations in illite crystallinity are apparent in various size fractions within a sample as a result of a relict detrital imprint in the coarsest fractions.

The general mineralogical composition of the clay size fraction comprises omnipresent illite/ muscovite, chlorite with several samples of diagenetic to low anchizonal level having kaolinite present. Discrete paragonite and mixed layer paragonite/muscovite occur preferentially in Devonian rocks of anchi- and epizonal grade. A distinct pyrophyllite bearing horizon, which indicates a local Al-enriched whole rock composition, can be traced from Combe Martin into the Quantock Hills.

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Introduction

The area of the "Culm synclinorium" and the adjacent Devonian rocks to the north have received little attention in terms of their regional metamorphic evolution. Previous work has focussed on the district between the Dartmoor and Bodmin Moor granites (Isaac, 1982) or has not extended further north than Bude (Phillips, 1928; Primmer, 1985; Robinson and Read, 1981) with the exception of a reconnaissance study of Brazier *et al.* (1979). A study relating to stratigraphic position and provenance of clays has been undertaken in the Crackington Formation by Grainger and Witte (1981) Summaries of the tectonic and sedimentological history of the northern half of the south-west England peninsula (e.g. Dearman, 1963, 1970; Simpson, 1971; Matthews, 1977; Sanderson, 1984) and in particular the tectonic zones introduced by Shackleton *et al.* (1982) suggest that significant changes in metamorphism can be expected. The present work is therefore to establish the metamorphic character of the formations ranging from the Lower Devonian Lynton Shale to the Westphalian Bude Formation.

A general outline of the geology is given in Matthews (1977) and Durrance and Laming (1982).

Methods and experimental procedure

A difficulty in the assessment of incipient metamorphism in terrains dominated by pelitic lithology is that most of the key mineral reactions which may be used to pinpoint temperatures are established for volcanic rocks or their immediate sedimentary derivates the greywackes. Owing to the absence of such units, the increase in crystallinity of illite with increasing metamorphism has been employed as a relative indicator of grade. Illite is used here in a general sense and covers the $2M_1$ polytype of illite and muscovite which is seen at higher grade.

The indices measured are:
1. The Kubler index [IC], (Kubler, 1967, 1968) which represents the width of the 001 illite peak (or 002 illite, muscovite peak for the $2M_1$ polytype) measured at half peak height. It decreases with increasing metamorphism. 2. The Weaver index (Weaver, 1960) is the sharpness ratio of the 001 illite peak measured at the 10Å/10.5Å positions.

For both indices the peak positions are measured minus the background level.

The boundaries used here are Δ 0.38 2θ [IC], 2.3 [Weaver] for the step from diagenesis to anchizone and Δ 0.21 $2\boldsymbol{\theta}$ [IC], 12.1 [Weaver] for the anchizone to epizone transition. The assignment of the Kubler index is a subject of constant debate, because of its dependence on experimental parameters (Kisch, 1980a,b).

In estimating the relative metamorphic grade a plot of IC. versus Weaver index has been used, but in borderline cases the IC is preferred due to its greater sensitivity in the anchi, and epizone fields.

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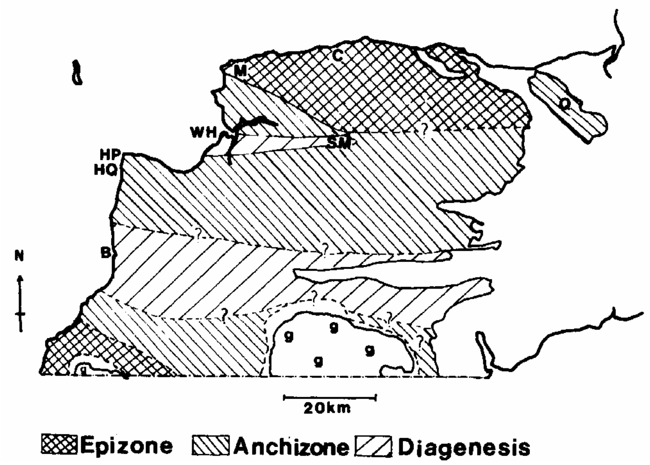

Fig. 1. Preliminary map of the metamorphic zones in North Devon and West Somerset (185 samples). The boundaries west of Dartmoor are analogical to T.J. Primmer (1985). Boundaries have been drawn where the smallest size fraction reaches the subsequent higher grade. B - Bude, C - Coombe Martin, HP - Hartland Point, HQ Hartland Quay, M - Morthoe, Q -Quantock Hills, WH - Westward Ho!

The IC and Weaver measurements have been determined using a Philips PW 1730 diffractometer with Ni-filtered, Cu-Kα radiation at a tube rating of 40 KV and 40 mA with an automatic divergence slit, a receiving slit of 0. 1mm and a scanning speed of 1/2º 2θ/min.

All samples have been separated into four size fractions using a centrifuge technique (> 1um <4um; >1um <2um; >0.5um < 1um; < 0,5um); for further experimental details see Hathaway (1958).

In the absence of suitable geobarometric mineral reactions, the empirical method proposed by Sassi and Scolari (1974), which is based on the bo dimension of the K - white micas, has been used. This spacing measured from the position of the 060 reflection is controlled in the first place by Fe- and Mg- substitution in the octahedral sheets and only to a minor extent by interlayer cations. It is employed as a measure of the phengite/celadonite content of a mica, which increases with increasing pressure within the temperature range comprising the upper anchi and epizone (Bertagnini and Franceschelli, 1982; Padan *et al.,* 1982). These authors also describe further constraints of the method depending on the mineralogical composition of the samples. This reflection also serves to distinguish between di- and trioctahedral micas.

Experimental parameters for bo determinations that differ from the IC and Weaver indices are a scanning speed of 1/4º 2θ/minute, over the range 58- 63º 2θ, using the quartz 211 peak (1.541Å) as an internal standard.

General metamorphic pattern (fig. 1)

The Upper Carboniferous rocks show a range of IC values from a mid-diagenetic to medium -anchizone level. Along the west coast a steady decrease in IC from Δ 0.78° 2θ (max.) in the Bude area to ca. Δ 0.3° 2θ south of Hartland is observed. The smallest size fraction value (< 0.5um) appears to "lag" in terms of its grade relative to the other size fractions.

From Hartland the metamorphism decreases again towards mid diagenetic crystallinities in the oldest strata of the Upper Carboniferous to the south-west of Westward Ho! (fig. 1).

The Devonian outcropping along the west coast reveals a steady increase of grade into the epizone, which is recorded in all four size fractions south of Morthoe.

Within the Devonian and Lower Carboniferous the increase in metamorphism roughly coincides with the change in stratigraphic level (fig. 2), though an occasional crosscut of the stratigraphic boundaries (e.g. the shallow upfold of the Baggy Sandstone south of Brayford SS686 346) is seen.

Such a coincidence between stratigraphic level and grade is not found in the Upper Carboniferous sequence, notably the metamorphic low near Westward Ho!.

The Devonian rocks of the Quantock Hills display upper anchizonal illite crystallinities (fig. 1).

Grain size effects on the metamorphic pattern (fig. 3)

A comparison of the IC values for the Upper Carbon-iferous samples from the coastal profiles (fig. 2) for all size fractions (e.g. of the < 0.5um and > 2um < 4um fractions in fig. 3) shows that for all fractions above 1um there are few values within the diagenetic field.

However in samples from the northern part of the profiles all size fractions show an overlap of values from mid anchizone into the epizone. This suggests that the coarser material in the Upper Carboniferous is influenced by relict detrital clays and that further north the phyllosilicates in all size fractions have undergone sufficient recrystallisation to lose the detrital affect.

An attempt to measure small scale variations in illite crystallinity within a fold (Hartland Quay SS 225 248) has not proved successful. Variations rarely exceeded two or three times the standard deviation (< 0.02º 2θ, n=30 for selected standard samples), and no systematic trend has been noted. This suggests that on such local scale the deformation has had no detectable systematic effect on the observed crystallinities; a conclusion suggested previously by Flehmig and Langheinrich (1974).

General mineralogy of the clay size fraction

The summary given in fig. 4 omits the coarsest size fraction, because of its possible clastic character.

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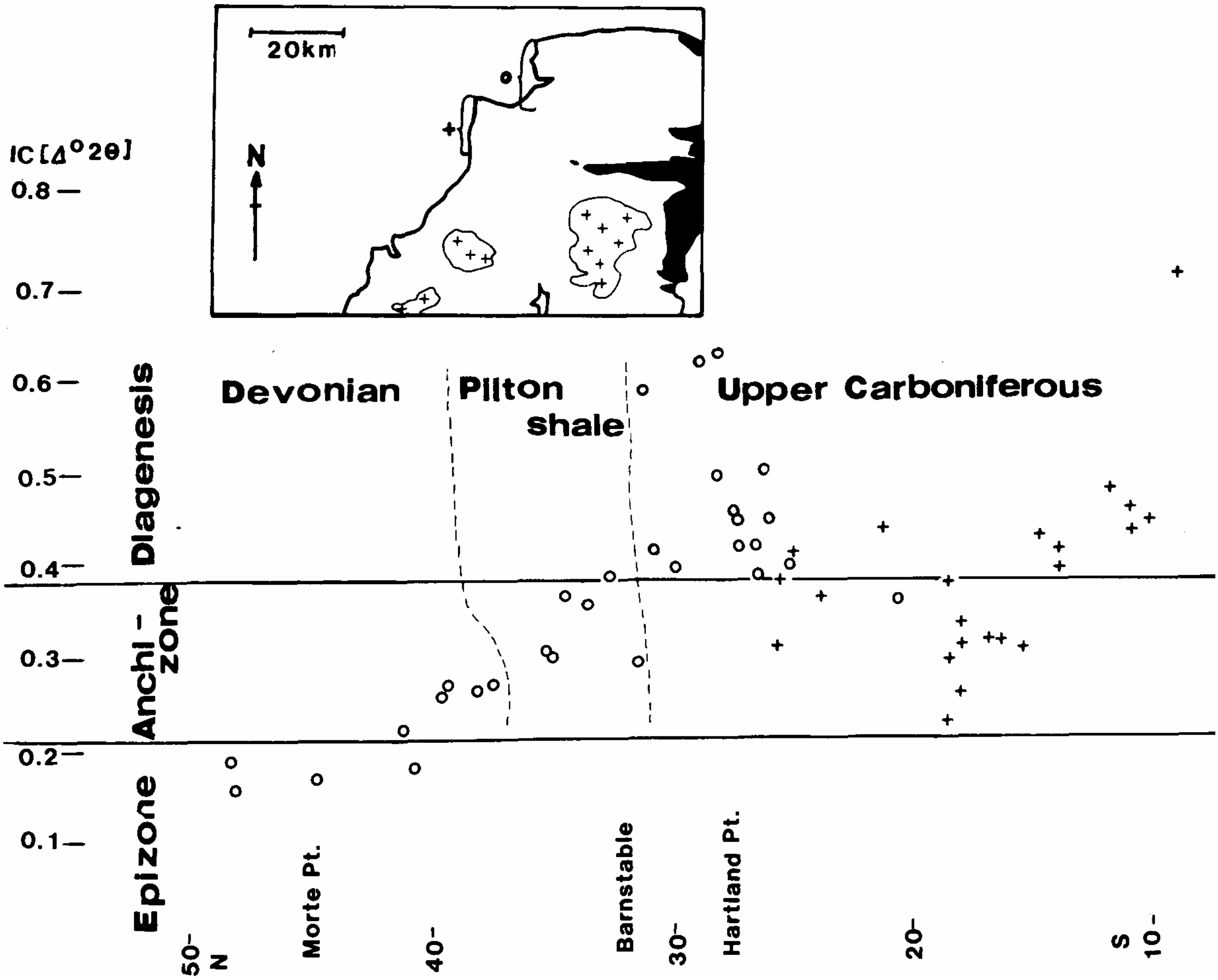

Figure 2. IC for the > 0.5 um < 1 um fraction of samples from the west coast plotted versus their geographical position in a north - south profile.

The illite of the diagenetic field frequently contains < 15% smectite interlayering, as determined using the method of Srodon (1984). However this method of smectite estimation has not proved fully satisfactory, as many samples plot outside the illite and illite/smectite boundaries given in fig. 2 of Srodon (1984).

Interlayering in the illite continues into the anchizone with <6% smectite. A decrease in the smectite content with increasing size fraction is evident in some samples but there is no definite trend. In the anchizone the 10Å phase is increasingly identified as the 2M1 polytype (Z87A), but in the lower part of this zone its presence may still reflect some clastic influence. In the epizone the $2M_1$ polytype is also recorded in the finest size fraction.

Paragonite interlayered with illite is first identified in samples from the upper part of the diagenetic field and is more frequently present in the anchizone.

Apart from its presence in the diagenetic field, kaolinite is observed in the lower anchizone in all four size fractions, but the occurrence at the anchi- to epizone boundary is confined to the pyrophyllite bearing, A1 - rich samples.

The main accessory minerals constitute a little feldspar in the coarser fractions plus quartz and calcite. The latter is generally confined to the smallest size fraction of shales. At the present preliminary stage of the study only two possible ideas about the origin of the calcite are outlined here.

1. The calcite is a metamorphic product. However if calcite is formed during progressive metamorphism in the range of mid diagenesis to epizone an increase in calcite grain size should be expected. It is therefore difficult to explain its restriction to the smallest size fraction in shales but an exception is seen in Fe - poor samples (see below).

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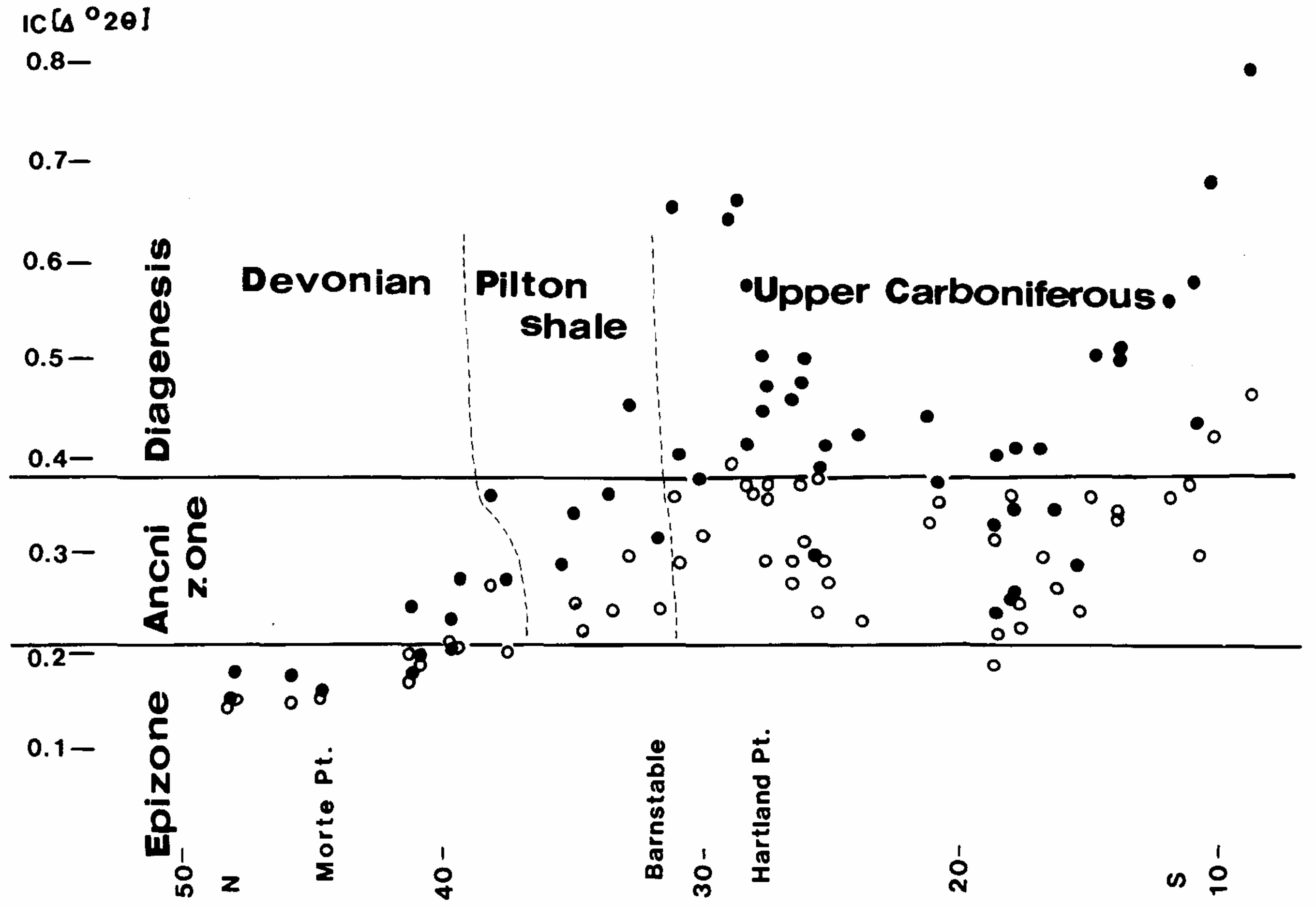

Figure 3. Plot as Fig. 2 for the <0.5 µm (dots) and the> 2 um <4 µm (circles) fractions. Further explanations see text

2. The calcite is preserved as a relict phase in finest fissures and voids in the shale. These are too small to participate in a larger scale fluid circulation system thus preventing it from being leached. A similar process has been described by Hower et al (I976) for diagenetic material, although these authors have not reported calcite in the < 0.5 um fraction.

Quartz in generally found in the > 1 um fraction, but samples from faultzones (e.g. Sticklepath SS 396 260) show the mineral already in the < 0.5 um fraction.

Iron - poor assemblages

These samples deviate from the general pattern by their complete lack or minimal content of chlorite. Beside illite/muscovite and discrete paragonite (identified by 9.7Å, 4.9Å 3.2Å peaks) quartz and calcite are also present. In addition some samples contain pyrophyllite and kaolinite. These Fe-poor assemblages so far have been detected in the Middle Devonian Wild Pear Slates, Little Hangman Grits and Avill Group (Webby, 1965) in the Brendon Hills and the corresponding strata in the north east Quantock Hills. A pyrophyllite bearing horizon which indicates local A 1- enriched whole rock composition can be traced from Combe Martin into the Quantock Hills (ST 207 345).

Fig. 5 represents microprobe analyses made of two pyrophyllite bearing samples displayed on a Velde diagram. Though designed for lower temperatures (Velde, 1983) this plot is preferred to the A1203, KAl02, NAA102 triangle used by Guidotti (1984), because the proportions of sodic and potassic feldspar have not been determined.

The points represent average compositions of all the phases present in the size fractions and no significant difference amongst them is apparent. The samples are from the upper anchi and epizone. Their homogeneous chemical character suggest that metamorphism has been sufficient to reach a chemical equilibrium for all size fractions examined.

bo - pressure indicator

bo - determinations are limited to samples from the Devonian outcrops because of the temperature constraints on the method (Padan *et al.* 1982), The mean value of 8.988 Å (s.d. = 0,0287, n=28) indicates very low pressure conditions and is comparable to the lowest pressure

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Bosost - facies (Pyrenees) described by Sassi and Scolari (1974).

Samples from the Lynton Shale and the Hangman Grits are found to be too rich in quartz to allow a satisfactory determination (Padan *et al.,* 1982; Sassi and Scolari, 1974). No regional variation is apparent from the results.

Conclusions

1. Temperature estimates:
The low smectite content of illite and the presence of paragonite - a possible user of sodium set free by progressing illitisation - plus the presence of chlorite in the small size fractions as well as in the coarse ones suggests temperatures of at least 100 - 150°C for the lowest grade Upper Carboniferous rocks. This estimate is based on a comparison with similar assemblages described from the Southern Appalachians by Weaver (1984).

The infrequent occurrence of kaolinite corroborates this conclusion, although variations in detrital origin are possible (Grainger and Witte, 1981).

Pyrophyllite is not useful for delimiting the temperatures of metamorphism. Its experimental formation from kaolinite and quartz at temperatures between 270°C and 300°C is described by Hemley et al. (1980) and Thompson (1970). This temperature can be substantially lowered to ca. 220°C, if a low aH20 is assumed (Frey, 1978).

The upper temperature limit for the region is set by the absence of biotite, which occurs in pelitic rocks at temperatures of 425 - 450°C (Winkler, 1979).

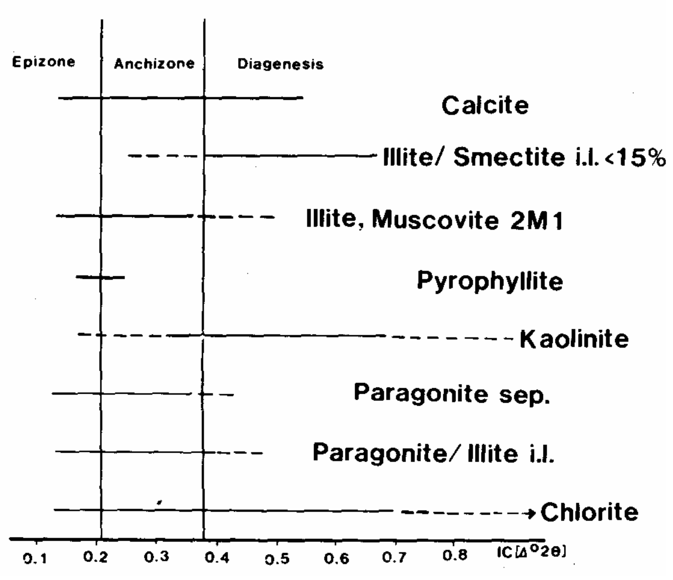

Figure 4. Summary of the mineralogy of the < 2 um fraction in the three metamorphic zones. Samples from all formations are included. ---: few samples available in that IC range.

2. Course of metamorphism:
For the Devonian and Lower Carboniferous rocks the increase in metamorphism with increasing age of the strata suggests an influence of burial metamorphism.

The very slightly altered rocks of the Crackington Formation near Westward Ho! comprise the oldest sections of the Upper Carboniferous, which appear to be one of the least deformed in the Culm synclinorium (De Raaf et al., 1965), Further south rocks equivalent or those younger in age show a more intense overall deformation (e.g. Hartland) and a higher metamorphic grade.

The area in the immediate southern vicinity of Westward Ho! may well represent a "frozen" state of burial metamorphism, whereas the more intense deformation elsewhere could have had some effect in leading to anchizonal metamorphism, allowing a further crystallisation of illite in the small size fractions but retaining a clastic overprint in the coarser. The relation of thrusting to increased metamorphism in the southernmost outcrops of the Crackington Formation (Ashburton area) has been noted by Grainger and Witte (1981).

The lower metamorphism in the Quantock Hills compared to the epizonal stage reached along the Bristol Channel coast further west could be explained by late offset along a wrench fault (described by Dearman, 1963).

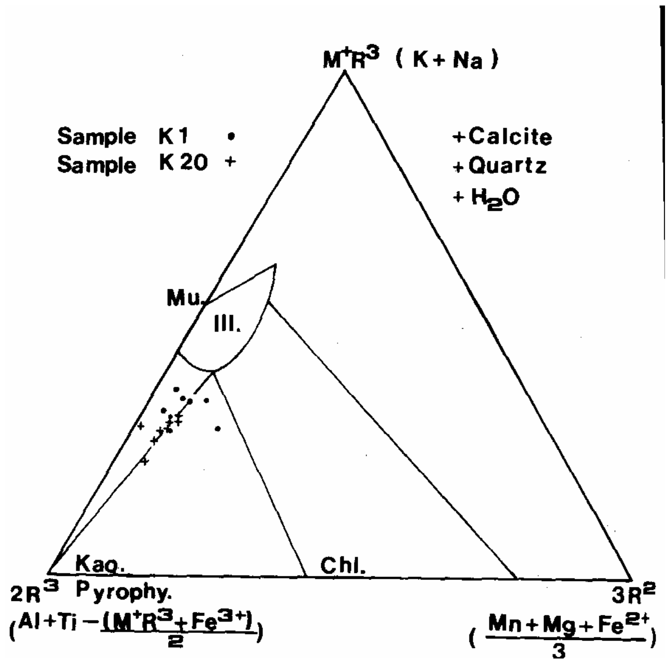

Figure 5. Microprobe analysis of two Pyrophyllite bearing samples for all size fractions (with defocused beam spot and varying diameter) represented on a Velde diagram. For detailed mineralogical composition see text. A JEOL JXS 733 (EDS) microprobe at Imperial College, London has been used.
Chl - chlorite, Ill - Illite, Kao - kaolinite, Mu - muscovite, Pyrophy - Pyrophyllite. Sample locations: K 1 ST 207 345, K20 SS 581 479.

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The Fort de la Latte Complex: late Pentevrian plutonism in Penthièvre, Brittany

MARY M. SHUFFLEBOTHAM*

Shufflebotham, Mary M. 1986. The Fort de la Latte Complex: late Pentevrian plutonism in Penthievre, Brittany. *Proceedings of the Ussher Society, 6,* 344-351.

The Fort de la Latte Complex forms the northern end of the Penthievre crystalline massif in N Brittany. It comprises foliated quartz diorites, which contain small enclaves of both cognate and accidental origin, and larger gabbroic bodies. All the rocks show evidence for at least a low-grade metamorphic modification.

It is suggested that the Fort de la Latte intrusion may have been emplaced during the same broad plutonic cycle which produced the late Pentevrian foliated granodiorites of the SE side of the Baie de St-Brieuc (Jospinet Group). Major and trace element geochemistry accord with a model of calc-alkaline plutonism in a convergent plate-tectonic regime.

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Introduction

The region of Penthièvre in N Brittany extends from St-Malo to St Quay - Portrieux, and lies in the Armorican Massif of NW France. As a result of the work of Cogné (1959), this region was considered to contain the type-area of Pentevrian (pre-Brioverian) outcrop. The Brioverian is a late Proterozoic sedimentary and volcanic sequence (review in Roach 1977), represented in the area of study by the Lower Brioverian Erquy, Hillion, and Fresnaye Volcanic Formations, and the Port a la Duc Formation (Fig. 1).

In 1959, at the locality of Jospinet (Fig. 1), Cogné described a major unconformity between the Lower Brioverian "Série d'Erquy' and the crystalline massif to the SE, which he assigned to the Pentevrian cycle. This basement is referred to here as the Penthièvre crystalline massif. Several other areas in the Armorican Massif were subsequently correlated with the Pentevrian. On the French mainland, these included parts of the St Quay - Portrieux area on the W side of the Bale de St-Brieuc (Ryan 1973); the St-Malo Migmatite Belt (Brown 1974); parts of the Cap de la Hague area in Normandy (Power 1974); and parts of the Trégor (Verdier 1968). In the Channel Islands, basement equated with the Pentevrian has been recognised on Guernsey (Adams 1967, 1976; Roach *et al.* 1972). Gneisses assigned to the Pentevrian have also been documented from Sark (Gibbons and Power 1975), and Alderney (Bishop *et al.* 1975).

Detailed mapping of the area from Pointe de la Latte to St Brieuc, and in particular, of the Penthièvre crystalline massif (Fig. 1), suggests a sequence of Precambrian events as follows:

1. the formation of an older complex of Pentevrian gneisses (the Yffiniac and Port Morvan Groups). These are basic to acidic in composition, with minor schists and amphibolites. They suffered polyphase deformation and metamorphism prior to 2;
2. the emplacement and subsequent deformation of Pentevrian granodioritic plutons (the Jospinet Group), and possibly the Fort de la Latte Complex;
3. a late Pentevrian shearing deformation which formed the Hillion - Cesson Mylonite Belt (Roach and Shufflebotham, unpublished data);
4. Brioverian sedimentation and volcanism;
5. deformation of both the Pentevrian and the Brioverian during the end-Precambrian Cadomian orogeny.

The Yffiniac Group contains agmatites, which were themselves agmatised by the gneisses of the Port Morvan Group. The polyphase deformation of the Yffiniac and Port Morvan Groups included phases of folding, as well as amphibolite facies metamorphism, prior to the emplacement of the Jospinet Group. The latter generally show only weak deformational fabrics. The dominant foliation in the Penthière crystalline massif is N-S to NE-SW, a trend described by Cogné (1972) as typical of Pentevrian massifs. In contrast, the Brioverian has E-W to ENE-WSW Oriented bedding and foliation in this area.

The granodiorites between Jospinet and Hillion (Jospinet Group) were shown to belong to the Pentevrian cycle by Cogné (1959); they underlie the base of the Brioverian at Jospinet, and pebbles and boulders derived from these granodiorites may be found in the basal Brioverian conglomerate at Jospinet, Hillion, and Cesson (Fig. 1). At Hillion and Cesson, the conglo-

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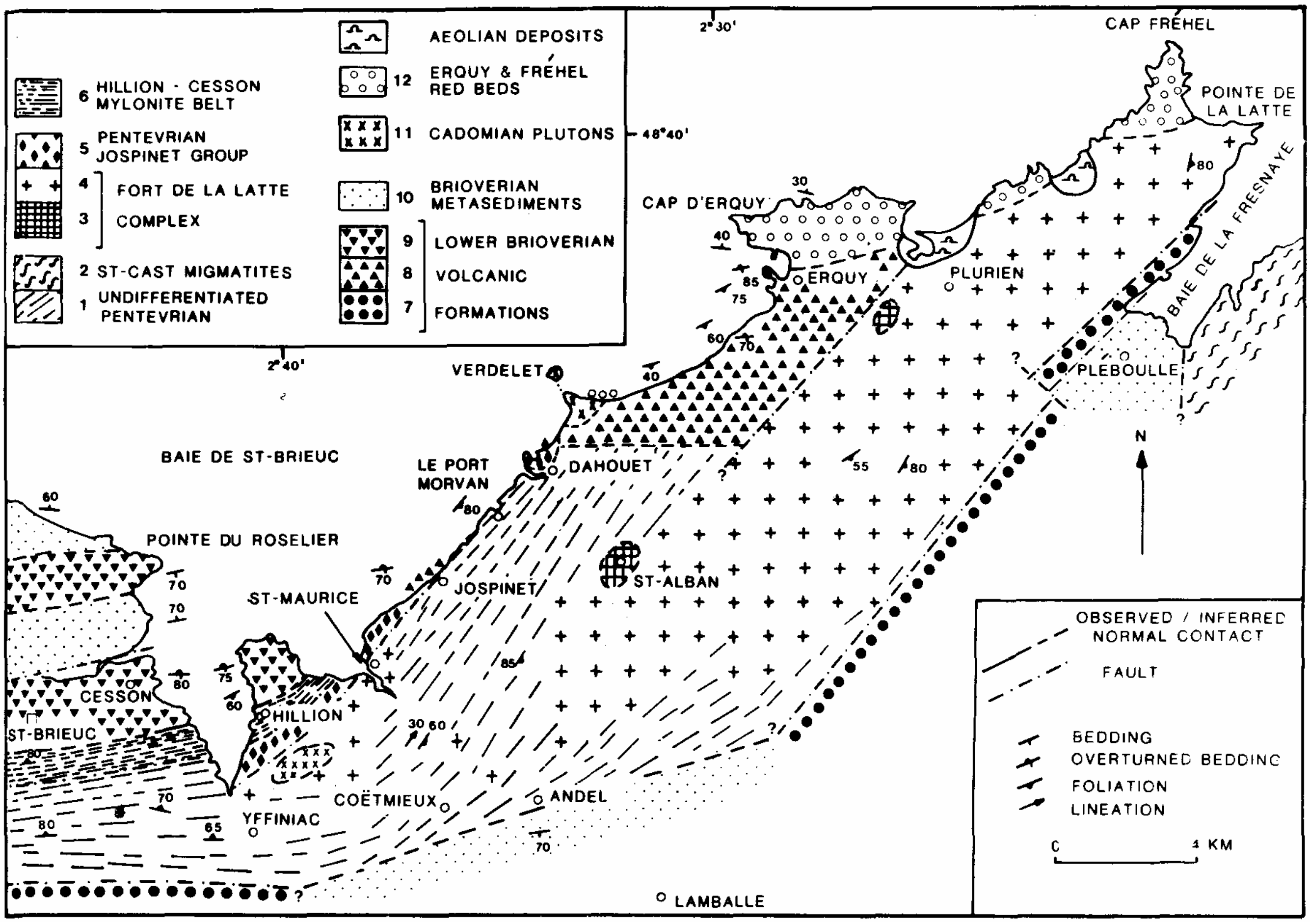

Figure 1. Simplified geological map of the Penthievre crystalline massif and its adjacent supracrustal formations.
I Pentevrian Yffiniac and Port Morvan Groups. 3 Gabbros. 4 Quartz diorites (including St-Maurice type). 7-9 Lower Brioverian volcanics: 7 - Fresnaye Volcanic Formation; 8 - Erquy Volcanic Formation; 9 - Hillion Volcanic Formation. 10 Port A la Duc Formation.

merates were deposited upon the eroded Hillion- Cesson Mylonite Belt. The mylonites are thought to have been derived from a heterogeneous assemblage of Pentevrian basement rocks.

The components of the basement on the flanks of the Baie de St-Brieuc are intruded by abundant metabasic sheets (Dahouet-type dykes), which cut across all Pentevrian structures. These sheets are interpreted as possible feeder dykes to the Lower Brioverian Erquy and Hillion Volcanic Formation (the Etage d'Erquy, Etage de Lanvollon, and Etage de Cesson - Locquirec of Cogné (1972)).

The Penthièvre crystalline massif has not, however, been generally accepted as Pentevrian. In 1974, Vidal et al. produced a Rb-Sr whole-rock isochron date of 631± 60 Ma, and a U-Pb zircon date of 593± 15 Ma on the intrusive Fort de la Latte foliated quartz diorite. (Rb-Sr dates quoted here have been recalculated using □Rb = 1.42 x 10-11 yr-l), These dates were interpreted as indicating the emplacement of the Fort de la Latte intrusion during the end-Proterozoic/early-Palaeozoic Cadomian orogeny. In 1971, Vidal *et al.* had obtained a date of 482 ± 10 Ma for the Erquy volcanic rocks. This was thought to indicate extrusion during Lower Ordovician times. However, Brown and Roach (1972a and b) and Shufflebotham and Roach (1985) maintained that the Erquy volcanics were Lower Brioverian, with the Ordovician date reflecting a metamorphic overprint. Cogné (1976) and Cogné et al. (1980) have assigned the rocks of the crystalline massif to the Cadomian cycle.

This paper is concerned with the relationship of the granodiorites of the Jospinet Group and the Fort de la Latte Complex.

Relationships of the Fort de la Latte Complex and the Pentevrian Jospinet Group.

The Fort de la Latte Complex forms the northern part of the Penthièvre crystalline massif; it chiefly comprises quartz diorite, with minor diorite, gabbro, and aplite. The complex is bounded to the west by the Erquy Volcanic Formation, and to the east by the Fresnaye Volcanic Formation and the turbidite-dominated Port à la Duc Formation (Fig, 1). Inland exposure in Penthièvre is very limited and contacts poorly defined. However, the

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western margin of the plutonic complex appears to be a fault, while the eastern contact can clearly be seen to be strongly tectonised on the NE side of Baie de la Fresnaye (Fig. 1). That there are neither sheets of metabasite in the quartz diorite, nor of quartz diorite in the metavolcanics, and the absence of contact-metamorphic effects, makes the relative ages of the rock types equivocal. Both the Erquy Volcanic Formation and the Fort de la Latte Complex are overlain unconformably by the Erquy and Fréhel Red Beds. These latter have been assigned various ages ranging between Cambrian and Permo-Carboniferous, though recently a Cambro-Ordovician age has been favoured (Auvray *et al.* 1980).

The Jospinet Group is faulted against the older gneisses, and is unconformably overlain by the Erquy Volcanic Formation at Jospinet (Fig. 1). The granodiorites range in composition from hornblende (now chlorite) - granodiorite, through biotite-granodiorite to a muscovite-bearing leuco-granodiorite. The intrusive sequence is unclear as boundaries between the separate intrusions are almost always faulted. However, leuco-granodiorite has been observed to cut biotite-granodiorite, suggesting a temporal relationship. Both these facies are cut by intermediate to basic porphyritic sheets, which also cut the Port Morvan Group. These sheets are in turn cut by Dahouet-type dykes.

At St-Maurice (Fig. 1), quartz diorites of the same type seen at Fort de la Latte are exposed, and are probably in faulted contact with the granodiorites of the Jospinet Group. Veins of granodiorite cut the quartz diorite. Dahouet-type dykes are abundant at St-Maurice, in contrast to the Pointe de la Latte area, where none have been identified.

Field relationships and petrography of the Fort de la Latte Complex

Quartz diorites of the Fort de la Latte Complex are well-exposed on the coast in the vicinity of Pointe de la Latte (Fig. 1). The southerly extent of the complex is not clear; it forms the major part of the massif as far S as Andel, while further SW, screens of quartz diorite can be identified within the Yff'iniac Group.

The main component of the Fort de la Latte Complex is foliated coarse quartz diorite. Its primary igneous mineralogy comprises essentially plagioclase (An_{34-39}) (45-60%), hornblende (<10-26%), and quartz (<10-25%). Biotite is usually present (up to 12%) but may be totally replaced. Up to 5% modal orthoclase can be present. Ore, apatite, sphene, and zircon are important accessories, often as inclusions in hornblende. The quartz diorite has a random to elongate fabric, dominated by weakly-aggregated plagioclase and hornblende. The plagioclase often shows deformation-twinning. Xenomorphic quartz, in aggregates up to 5mm long, shows undulose extinction, and in high-strain areas, has recrystallised to a micro-granular aggregate. The secondary mineral assemblage consists of prehnite, chlorite, epidote, and sphene. Prehnite is seen almost exclusively as a replacement in biotite, in which it forms spindle-shaped "bows" up to 2mm long. With chlorite and epidote, it may totally replace biotite, and like the biotite, can be kinked.

The secondary assemblages are considered to be due to a low-grade regional event rather than to retrogression accompanying cooling and uplift. No systematic change in grade is evident around Pointe de la Latte. The assemblages here indicate rather a partial metamorphic overprint within the prehnite-pumpellyite facies. Southwards towards Andel, the biotite is a reddish-brown metamorphic variety, often with secondary sphene grown along its cleavages. Partial replacement of hornblende by actinolite is evident, but prehnite has now become unstable. This apparent increase in grade to the SW in the Fort de la Latte Complex corresponds to that observed in the metasupracrustal rocks of the Erquy and Hillion Volcanic Formations. Along the SE coast of the Baie de St-Brieuc, the metamorphic grade of this Brioverian sequence is seen to increase from prehnite-pumpellyite facies at Erquy, to highest greenschist facies at Cesson (Fig.1).

The quartz diorite displays a fabric orientated at 0-40°, dipping steeply to the SE. This consistent fabric, and the parallelism of igneous hornblendes, suggest the intrusion was emplaced syn-kinematically. The dominant fabric in the Fort de la Latte Complex is parallel to the N-S to NE-SW trend characteristic of the older Pentevrian parts of the massif. It should be noted that a primary fabric is absent in the granodiorites of the Jospinet Group. These contain a cataclastic fabric which is considered to have formed during the late Pentevrian shearing event which produced the Hillion - Cesson Mylonite Belt.

Contained in the quartz diorite are numerous dark enclaves of both accidental and cognate origin. The accidental inclusions, which may reach 10m in length, were possibly derived from the Yffiniac Group; they show evidence for folding and amphibolite-facies metamorphism prior to their inclusion in the quartz diorite. They are generally composed of plagioclase and hornblende in an equant to elongate fabric, with minor quartz, chlorite, and epidote. Cognate inclusions are ubiquitous, though individually of much smaller size (generally < lm in length, often < 0.2m). These enclaves show strong flattening within the plane of the foliation of the host, supporting a syn-kinematic origin for the intrusion. These inclusions are of dioritic composition, consisting of plagioclase (An_{41-44}) and hornblende. Biotite is variable in amount, and often altered to epidote, prehnite, and chlorite. The cognate inclusions show random fabrics in their cores, but are slightly foliated on their margins. The petrographical similarities between these and the host suggests that they formed from the melt during crystallisation, rather than being restites of

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the source magma. It is envisaged that these inclusions began to crystallise in a relatively stress-free environment, but as the host solidified around them, under compression, they were strongly flattened.

Cutting the enclaves and quartz diorite are several aplitic sheets, of up to 0.5m thickness. These can show continuation of the fabric through them, reflecting emplacement before flattening of the complex ceased, or can be strongly foliated parallel to their margins.

A number of gabbroic lenses have been identified within the main body of the Fort de la Latte Complex, e.g. at the Barrage de Montafilan, some 2km from Plurien, and at St-Alban (Fig. 1). These gabbros are in a close, but obscure, relationship with the quartz diorite: they are apparently veined by it, and share a common foliation, but nowhere has any certain intrusive relationship been seen. They display a variety of mineral assemblages, with the primary mineralogy comprising plagioclase, hornblende, and clinopyroxene, and a secondary overprint reflected by several of the following: actinolite, sphene, chlorite, prehnite, epidote, zoisite, quartz, white mica. Partial replacement of clinopyroxene by secondary amphibole is common. Plagioclase (An_{48-74}) may be severely saussuritised, and occasionally recrystallised to more sodic plagioclase in small crystals or on grain margins. The plagioclase often shows deformation twinning. Whilst the precise relationship between the gabbros and quartz diorites is unclear, the gabbros may be viewed broadly as precursor intrusions to the main quartz dioritic mass.

The St-Cast Migmatite Belt is exposed some 4 km from Pointe de la Latte on the SE side of Baie de la Fresnaye (Fig. 1). The migmatites are poly-deformed, and cut across by a number of quartz dioritic dykes, of identical appearance to the Fort de la Latte quartz diorite. These sheets have never been seen in the Brioverian Fresnaye Volcanic Formation and Port a la Duc Formation. Brown (1974) assigned the St-Cast migmatites a Pentevrian age. However, this was refuted by Brun and Martin (1979), and Peucat (in press), who suggest a Brioverian age for the metasediments, with all the migmatisation being Cadomian.

As well as the syn-intrusive fabric present in the rocks of the Fort de la Latte Complex, a later deformational event is recorded by the network of zones of cataclasis and shear which cuts the quartz diorite and its enclaves. In single or conjugate arrays, with each zone up to 0.15m thick, they are typified by grain-size reduction and colour darkening.

Porphyroclasts of plagioclase in a quartz-chlorite-epidote pseudomatrix are common. These zones of movement show both sinistral and dextral movement sense, with no apparent preferred orientation. Such a fabric is thought to have resulted from a semi-brittle deformational event which post-dated the emplacement and consolidation of all parts of the complex. A similar fabric affects the granodiorites and St-Maurice quartz diorite of the SE flank of the Baie de St-Brieuc.

The deformation and accompanying metamorphism of the cover sequences have been assigned to the Cadomian (Cogné 1959; Brown and Roach 1972 a and b; Shufflebotham and Roach 1985). Brown and Roach *(op. cit.)* suggested that the supracrustal rocks at Erquy could not be Ordovician, as stated by Vidal *et al.* (1971) and Cogné *et al.* (1980), since they are metamorphosed. No evidence in support of an end-Caledonian event is recorded in this part of Brittany. The NE part of the Armorican Massif is thought to have escaped severe reworking during the Hercynian event (Roach, pers. comm.). Indeed, the only manifestations of the Hercynian in Penthièvre are the NW-SE - trending dolerite dykes which cut the late shear fabric in the Fort de la Latte Complex. The metamorphism and deformation of the complex are assigned to the same Cadomian event which deformed and metamorphosed the Briovenan.

Geochemistry

As part of a geochemical survey of Penthièvre massif, 45 samples from the Jospinet Group and Fort de la Latte Complex (including the quartz diorite at St-Maurice) have been analysed. Major and trace element determinations were made. on a Siemens SRS 200 X.R.F. spectrometer at the University of Bristol, and an ARL 8420 X.R.F. spectrometer at the University of Keele. FeO was determined by titration. Selected samples were

	1	**2**	**3**	**4**	**5**	**6**	**7**
SiO_2	49.29	50.01	58.63	54.82	64.90	68.32	72.78
TiO_2	0.23	1.05	0.74	0.91	0.56	0.24	0.10
$A1_2O_3$	19.47	20.05	17.03	18.33	15.33	17.30	14.77
Fe_2O_3	0.78	2.29	1.17	1.53	1.79	0.82	0.47
FeO	3.71	6.63	4.57	4.54	1.47	0.50	0.18
MnO	0.09	0.15	0.10	0.12	0.06	-	-
MgO	8.14	4.96	3.57	4.16	1.63	0.70	0.47
CaO	14.06	8.11	5.92	7.44	.3.29	6.94	0.50
Na_2O	1.71	3.58	3.78	4.83	4.22	6.62	6.00
K_2O	0.26	1.39	1.42	0.78	2.59	2.39	3.19
P_2O_5	-	0.13	0.15	0.24	0.13	0.07	0.04
H_2O	2.00	1.81	2.21	2.24	4.21	1.17	0.80
Total	99.74	100.14	99.29	99.93	100.18	99.08	99.30
Cr	496		54	72	60	8	10
Ni	110		20	57	23		
Nb		6	7	8	2	4	9
Y	5	19	22	22	11		1
Zr	19	84	122	153	103	117	76

Table 1. Major element and selected trace element data for typical rocks: **Fort de la Latte Complex -** 1 Gabbro 2 Cognate inclusion 3 Quartz diorite; **St-Maurice** - 4 Quartz diorite; **Jospinet Group -** 5 Hornblende-granodiorite 6 Biotite-granodiorite 7 Leuco-granodiorite.
Major elements in wt.% oxides; trace elements in ppm. * Below detection level.

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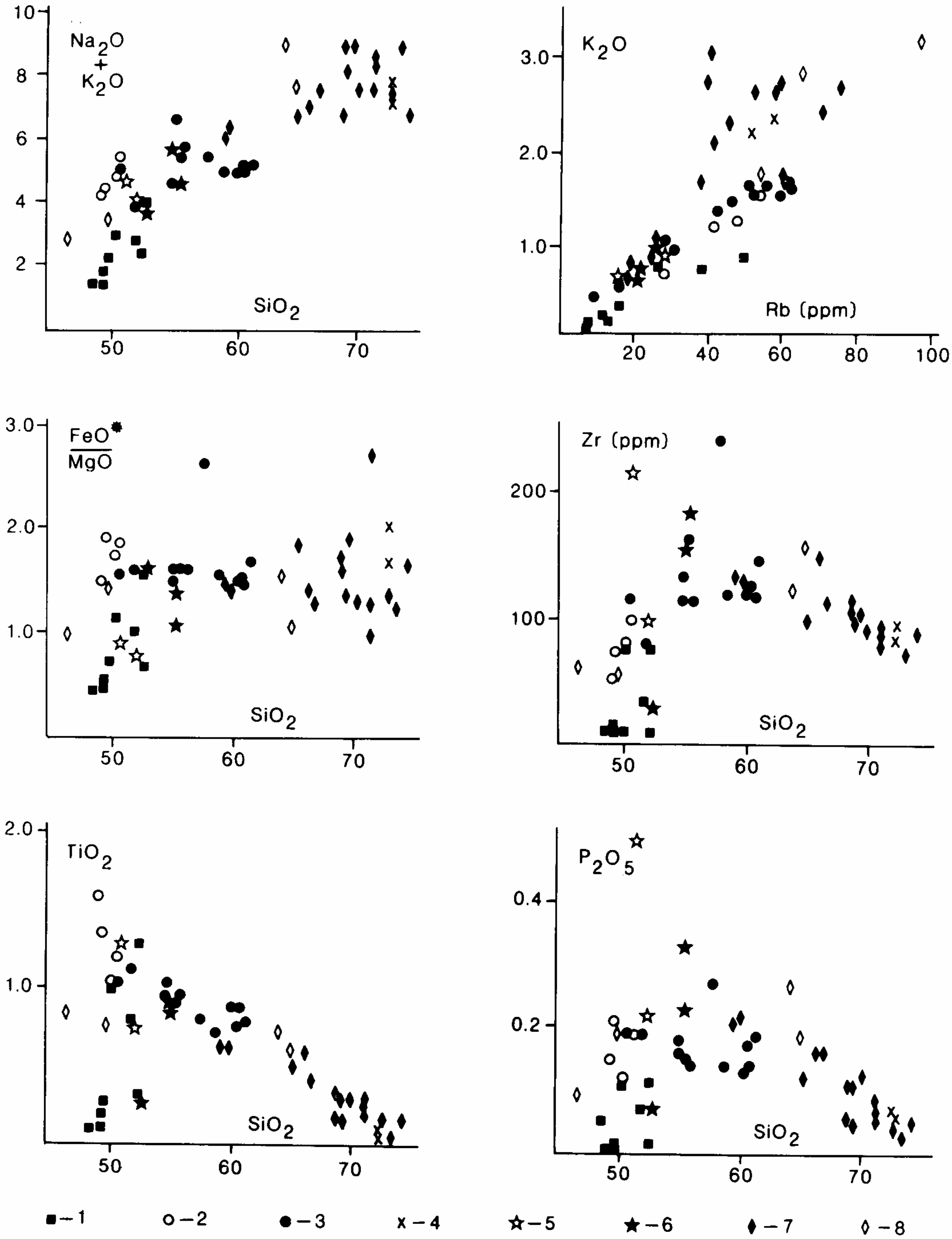

Figure 2. Plots of TiO_2, P_2O_5 $Na_2O + K_2O$ (in weight % oxides) against SiO_2; FeO*/MgO (as ratio of weight % oxides; FeO* = total iron as FeO) against SiO_2; Zr (in parts per million) against SiO_2; and K_2O (weight % oxide) against Rb (parts per million). Key: 1 Gabbros in Fort de la Latte Complex; 2 Cognate inclusions in Fort de la Latte quartz diorite; 3 Fort de la Latte quartz diorite; 4 Aplites in Fort de la Latte quartz diorite; 5 Inclusions in St-Maurice quartz diorite; 6 St-Maurice quartz diorite; 7 Jospinet Group granodiorites; 8 Basic and intermediate sheets in Jospinet Group.

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analysed for rare-earth elements by I.C.P. (Walsh *et al.* 1981) at King's College, London, and for incompatible elements at U.R.R., Risley, Warrington.

All these rocks are considered on structural grounds to be younger than the orthogneisses of the Yffiniac and Port Morvan Groups, but are cut, with the exception of the rocks in the N part of Penthièvre, by the Dahouet-type metabasic sheets.

A number of variation diagrams are shown in Fig. 2, and a representative set of analyses in Table 1. A wide range of SiO_2 values is observed, from 48 to 73%. With the exception of the alkalis, all the major and most trace elements show negative correlation with SiO_2. The gabbroic rocks tend to show some spread, with a few clustering together, and the remainder scattered. The cluster at the lowest SiO_2 end are those with the highest Cr and NJ. The gabbros are spatially related to the quartz diorites, but cannot be related simply by fractional crystallisation.

Certain petrographic features of the gabbros suggest that they might be, in part, cumulates. This is supported by the high Cr and Ni values for some gabbros (for the gabbro group, Cr = 63-622 ppm; Ni = 11-171 ppm), relative to the generally lower Cr and Ni contents of the quartz diorites and granodiorites (see Table 1).

Plots of TiO_2, P_2O_5, and Zr against SiO_2 are shown in Fig. 2. These display a small rise in the abundance of these incompatible elements with increasing SiO_2 until a SiO_2 content of about *55%* is reached, after which a steady decrease towards the SiO_2-rich end is observed. This behaviour suggests that Ti, P and Zr were acting compatibly at intermediate SiO_2 levels, and is reflected petrographically: sphene, apatite, and zircon are important accessory minerals in the quartz diorites.

The LIL elements,, Ba, Rb, and Sr, produce widely scattered plots (not shown), which reflect their mobility during alteration. However, a plot of K_2O against Rb (Fig. 2) shows a suggestion of two trends diverging from the origin, an indication that perhaps the gabbro - quartz diorite suite and the granodiorites evolved along different crystallisation paths, possibly from two sources.

Chondrite-normalised REE profiles for selected samples from the Fort de la Latte Complex and the Jospinet Group granodiorites are shown in Fig. 3. Included for comparison with the granodiorites are analyses from a granodioritic boulder in the Brioverian basal conglomerate and a mylonite from the Hillion - Cesson Mylonite Belt. The gabbro profiles show no fractionation of the REE (CeN/YbN = 1.1). The quartz diorites show fractionated profiles (average CeN/YbN = 5.09), with moderately strong LREE fractionation, and unfractionated HREE. The pattern mimics that of the cognate inclusion, but a small positive Eu anomaly in the latter suggests plagioclase accumulation, and is matched by a

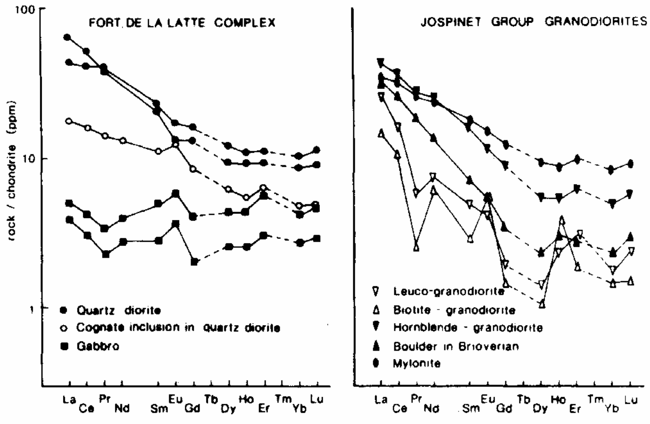

Figure 3. Chondrite-normalized rare-earth element profiles for the Fort de la Latte Complex and the Jospinet Group (normalisation factors of Nakamura 1974).

small dip in Eu in the host, in which mafic fractionation was probably a more important process. Such a combination of unfractionated REE patterns in gabbros and fractionated patterns in the intermediate intrusions has been noted in the calc-alkaline suites of the Coastal Batholith of Peru (Atherton *et al.* 1979). The REE data of the Jospinet Group granodiorites show a reverse to the normal igneous trend for such intermediate to acidic rocks. Here the more acidic components (leuco- and biotite-granodiorites) show the lowest total REE contents. Such anomalies have been discussed by Pankhurst (1979), who suggested that systematic depletion in HREEs in granitoids relative to intermediate

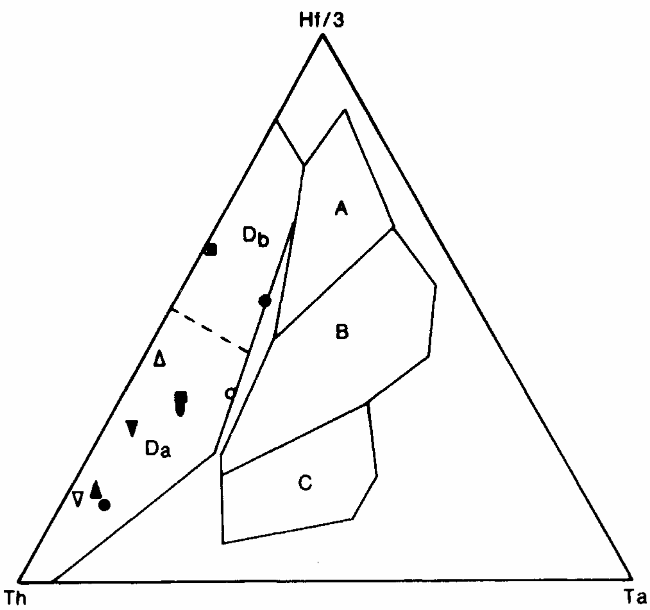

Figure 4. Th-Hf-Ta diagram of Wood (1980) for the same samples as in Figure 3. Symbols as for Figure 3. Fields as follows:
A: N-Type MORB; B: E-type MORB and tholeiitic within-plate basalts and differentiates; C: Alkaline within-plate basalts and differentiates; D: Destructive plate-margin basalts and differentiates-Db: arc-tholeiites; Da: calc-alkaline lavas.

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rocks, might be explained by progressive removal of amphibole by fractional crystallisation. Alternatively, there may be no simple relationship between the more acidic and intermediate members of the Jospinet Group.

All the rock-suites examined show calc-alkaline affinities. This is reflected in their limited Fe-enrichment; low Cr, Ni and TiO_2 contents in many of the intermediate rocks; and an abundance of hornblende- and biotite- rich lithologies, On an AFM diagram (not shown), the majority of samples plot well within the calc-alkaline field. Diagrams such as those of Pearce *et al.* (1984) (e.g. Rb v. Y+Nb and Y v. Nb) discriminate between different tectonic settings for granitoids. All the Penthièvre samples lie in the "volcanic-arc" or "syn-collisional" granite fields on such plots (not shown). Figure 4 shows a Th-Hf-Ta diagram (Wood 1980). This plot attempts to distinguish among tectonomagmatic settings of acidic to basic rocks. Selected samples analysed for Th, Hf and Ta all plot in Field D, that is, the field of destructive plate margin basalts and their differentiates.

Isotopic evidence

No new isotopic evidence has been obtained. The Fort de la Latte Complex was dated by Vidal *et al.* (1974) using Rb-Sr whole-rock isochron, U-Pb on zircons, and K-Ar and Rb-Sr mineral methods.

The Rb-Sr whole-rock isochron date of 631±60 Ma was derived from 7 samples. These have low Rb/Sr ratios (<0.135) and only a small range of 87Rb/16Sr values (0.087 - 0.390), suggesting they are not ideal for this type of isotopic work. It is possible to reset Rb-Sr whole-rock systems during a low-grade or greenschist facies metamorphism (Field and Räheim 1980) so the isochron may not in fact represent the age of emplacement. However, the Rb-Sr age is supported by a U-Pb date of 593± 15 Ma which was obtained on zircons from the northern end of the Fort de la Latte Complex.

The mineral dates obtained by Vidal *et al. (op. cit.)* on hornblendes and biotites from the north and south of the crystalline massif range between 590 and 490 Ma. They were interpreted by the authors as reflecting the influence of the emplacement of the Cadomian Granite d'Yffiniac (dated at 507 ±30 Ma by Vidal (1980)). However, this intrusion occupies only a very small area (c.1km diameter) at outcrop, and does not appear to have an extensive aureole indicating its presence at depth. Possibly a more realistic interpretation is that the mineral dates reflect re-opening of the Rb-Sr and K-Ar mineral systems during the Cadomian orogeny.

Discussion and conclusions

Field evidence suggests that there are structural similarities between the Fort de la Latte Complex and the Pentevrian granodiorites of the Jospinet Group. All the rocks show a N-S to NE-SW foliation which contrasts with the E-W to ENE-WSW trend developed in the Brioverian sequence. There is a cataclastic style of deformation and a low-grade metamorphic imprint in both rock groups, which are attributed to the Cadomian event.

The St-Cast Migmatite Belt is cut by sheets of quartz diorite, identical to that at Fort de la Latte. If the migmatite belt is of Pentevrian age, as suggested by Brown (1974), then the absence of the sheets in the adjacent Brioverian formations might be used as evidence for a late Pentevrian age for the quartz diorite suite.

The Fort de la Latte Complex and the Jospinet Group can be incorporated into a model for late Pentevrian calc-alkaline plutonism. The crystalline rocks have the characteristics of an I-type series (nomenclature of Chappell and White 1974), ranging widely in composition from gabbroic to aplo-granodioritic. The nature of the components accords with emplacement in a tectonic regime involving subduction. The change from calcic gabbro production, through true calc-alkaline plutonism, to peralkaline granodiorites, suggests increasing arc maturity with time (Brown 1982). Nystr6m (1982) has suggested that the Andino-type plutonism documented for the western American cordilleras during Mesozoic - Caenozoic time, was a feature of the Precambrian, and recognised I-type collisional intrusions in the post-Svecokarelian Proterozoic of Sweden. In Penthièvre, the absence of large quantities of the andesitic or dacitic volcanics normally associated with Calc-alkaline plutonism, presents a problem. Some clasts of intermediate and acidic volcanics have been noted in the heterolithic basal Brioverian conglomerate, while a few basic and intermediate sheets cut the granodiorites of the Jospinet Group. It is possible that the late Pentevrian intrusions represent the lower parts of a calc-alkaline arc, from' which the volcanic components had been eroded prior to Brioverian sedimentation and volcanism.

The following conclusions can be drawn:

1. Field evidence suggests that the Fort de la Latte Complex is late Pentevrian, and not Cadomian as suggested by Vidal, *et al.* (1974).
2. The Fort de la Latte Complex and the Jospinet Group represent an episode of late Pentevrian plutonism. Together with the older Pentevrian orthogneisses (Yffiniac and Port Morvan Groups), these intrusions formed the upper part of the continental crust through which the Brioverian volcanics were erupted.

The implication of the c.600 Ma age for the Fort de la Latte Complex is that Brioverian sedimentation occurred in a much shorter time span than previously realised.

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Read at the Annual Conference of the Ussher Society, January 1986

Geochemical features of Permian rift volcanism - A comparison of Cornubian and Oslo basic volcanics.

S.C. GRIMMER
P.A. FLOYD

Grimmer, S.C. and Floyd, P.A. 1986. Geochemical features of Permian rift volcanism - A comparison of Cornubian and Oslo basic volcanics. *Proceedings of the Ussher Society 6,* 352-359.

The post-orogenic Permian Exeter Volcanic Series (EVS) of Cornubia is volumetrically insignificant relative to the thick lava piles within the Oslo graben of similar age. Both sequences are dominantly basic in character and characterised by incompatible element enriched continental-type basalts with alkaline affinities. The Oslo volcanics generally have much higher absolute abundances of incompatibles, but lower LIL/HFS ratios and greater L/HREE fractionation. They also exhibit a number of geographically separate chemical suites, whereas the EVS basalts appear to form a single comagmatic fractionated suite. Overall patterns of LREE-enrichment in the basalts from both regions imply the influence of residual garnet in the source. One significant difference between the regions is the overall higher and variable Th/Ta and La/Nb ratios in the EVS basalts which may indicate a crustal contribution to these lavas, rather than specific mantle enrichment.

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Introduction

Permian volcanism in Cornubia (the Exeter Volcanic Series) represents a rather limited and localised eruptive episode compared with the far more extensive and temporally analogous volcanism of the Oslo graben continental rift in southern Norway, (Fig. 1).

Volcanic rocks from both regions have similar ages based on radiometric dating. The original K/Ar date(Miller and Mohr 1964) for the Cornubian basalts has been recalculated to 291±6 Ma (Thorpe *et al.* 1986) and places them in the lowest Permian. A Rb/Sr date of 292±8 Ma (Sundvoll 1978) for the earliest members of the temporally more extensive Oslo basalts falls below the Permo-Carboniferous boundary. The Cornubian volcanics lie concordantly within the lower part of the Permian sequence of continental sandstones and conglomerates (Laming 1966).

The Cornubian basalts show close geographical association with a series of incompatible element enriched lamprophyres which were erupted under similar conditions within small-scale fault bounded post-orogenic troughs or grabens (Whittaker 1974). Whereas the Cornubian volcanics take the form of minor local flows and vents, the Oslo basalts are characterised by widespread fissure-type eruptions and extensive shield volcanoes forming significant stratigraphic sequences (Segalstad 1978). The Oslo basalts in addition form part of a bimodal suite, the dominant member of which is the later acidic latite-trachyte group of "rhomb-porphyries'. In many cases the Oslo volcanics are also directly associated with large intermediate to acid plutons, an association not necessarily found between the Cornubian batholith and the volcanics of this area. One final significant difference is that whereas the Oslo graben is sited in and underlain by strongly attenuated early Proterozoic crust (Welin and Gorbatschev 1976) the pre-Permian basement of Cornubia was probably significantly thicker and late Proterozoic in age (Davies

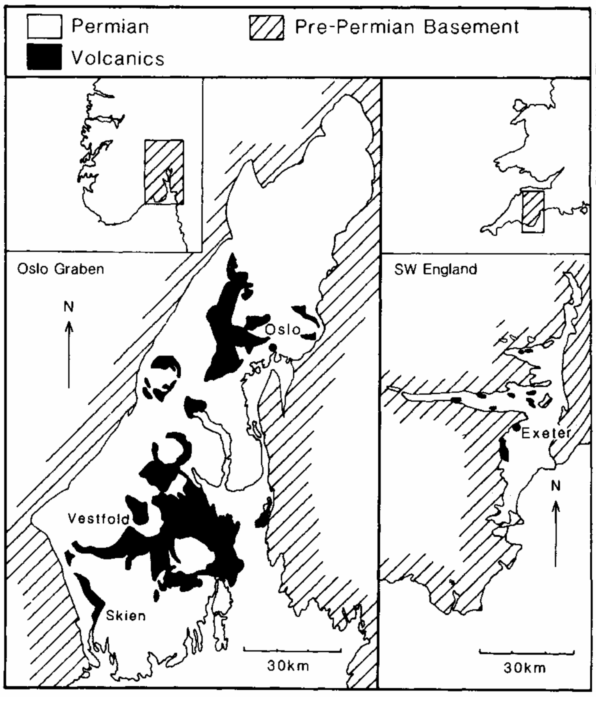

Figure 1. Comparative geological maps showing the distribution of Permian volcanics in the Oslo graben and SW England.

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1981). The post-orogenic setting of the Tiverton trough implies far greater crustal thickness during the early Permian.

Petrography

The Exeter Volcanic Series has been comprehensively described by Knill (1969), and can be broadly divided into two compositional groups. The earlier basaltic group is characterised by plagioclase and olivine phenocrysts, the latter now invariably pseudomorphed by Fe oxides. In addition clinopyroxene is found as small crystals in the groundmass with ubiquitous Fe-Ti oxides. Thorpe *et al.* (1986), report the presence of alkali feldspar, both as a mantling to the basic plagioclase and in the groundmass. The lavas are generally vesicular or amygdaloidal and commonly porphyritic although trachytic textures are found. The second group consisting of lamprophyres is more varied in nature, ranging from olivine minettes to mafic syenites (Knill 1969). Thorpe *et al.* (1986), describe their mineralogy as dominated by olivine and biotite phenocrysts with subsidiary clinopyroxene and analcite (occasionally after leucite) in an orthoclase-rich groundmass. All the lavas have been extensively altered by contemporaneous surface weathering and subsequent hydrothermal activity. Primary mafic phenocrysts are uncommon, being largely pseudomorphed by Fe-Ti oxides and smectite clays. Carbonate and zeolites also occur as secondary phases.

The basaltic rocks of the Oslo graben are also characterised by considerable variation in petrographic types. Due to the more extensive nature of the volcanism, direct stratigraphic correlations between individual members are possible. The earliest manifestation of Oslo graben volcanism occurs at Skien in the south of the area (Fig. 1) and is dominated by a thick sequence of ankaramitic basalts. The major crystallising phase is a titaniferous clinopyroxene found as large phenocrysts in a fine grained groundmass of clinopyroxene, Fe oxides and plagioclase. Occasional high-pressure phenocrysts of olivine, with mantles partially altered to actinolite, are also found. The ankaramites are overlain by later basanites with a primary assemblage of strongly zoned clinopyroxene and basic plagioclase phenocrysts set in a coarser groundmass of plagioclase microlites, clinopyroxene and ore minerals. Nepheline is occasionally found, and the presence of melilite has been reported by Segalstad (1976). At Vestfold in the central area of the graben (Fig. 1) is a reduced lava sequence dominated by alkali olivine basalts. Clinopyroxene and basic plagioclase comprise the major phenocryst phases, with subordinate amounts of olivine, now extensively altered to Fe oxide and iddingsite. The fine grained groundmass consists of clinopyroxene, feldspar and ore minerals. In both areas the lavas are variably amygdaloidal and commonly porphyritic in texture, although trachytic textures are found in certain of the Vestfold flows. As in the Cornubian volcanics hydrothermal alteration has been extensive. Common secondary phases are epidote, chlorite, albite and carbonate although in certain circumstances actinolite, prehnite and pumpellyite are found, indicating an overall higher degree of alteration up to greenschist facies metamorphism.

	S120	**S140**	**S147**	**V106**	**Vl19**
SiO_2	42.87	41.04	45.44	46.07	46.45
TiO_2	3.49	3.15	2.65	3.36	3.04
Al_2O_3	13.10	16.76	13.93	9.89	12.53
Fe_2O_3	14.23	12.71	12.03	11.38	13.02
MgO	7.16	6.07	6.58	7.46	5.32
CaO	9.26	9.24	10.82	9.23	8.32
Na_2O	3.33	2.83	3.07	1.87	2.59
K_2O	2.08	1.21	1.96	1.60	3.47
P_2O_5	0.48	0.83	0.45	0.49	0.51
Loss	2.34	4.70	1.30	8.23	2.27
Total	98.52	98.71	98.38	99.75	97.67
La	54.33	36.41	52.39	50.67	50.56
Ce	127.15	81.62	120.62	120.92	120.76
Pr	14.37	9.58	13.37	13.76	13.84
Nd	60.87	40.77	56.08	57.85	58.62
Sm	10.08	7.37	9.17	9.91	10.40
Eu	3.01	2.48	2.64	2.86	2.89
Gd	7.30	6.12	6.58	7.27	7.90
Dy	4.62	4.55	4.21	4.99	5.77
Ho	1.08	1.17	0.88	1.14	1.33
Er	2.18	2.40	2.10	2.58	2.98
Yb	1.30	1.65	1.32	1.69	2.13
Lu	0.19	0.27	0.21	0.27	0.33
Cs	3.49	0.33	1.58	1.41	1.21
Rb	67	23	60	33	96
Sr	1473	1680	1073	760	1460
Th	5.40	2.29	5.47	7.50	5.86
U	0.96	0.58	1.29	2.26	1.07
Nb	65	35	54	60	51
Ta	4.00	2.03	3.55	3.72	3.19
Zr	247	141	239	287	304
Hf	7.10	3.90	6.80	8.90	8.50
Y	23	27	21	25	32
Cr	148	0	95	337	122
Ni	114	59	93	156	68

Major elements by XRF, traces by XRF, ICP & INAA. Oxides in wt.%, traces in ppm, S120, Ankaramite; S140, Dolerite dyke; S 147, Basanite; V106 & V119, Olivine basalts.

Geochemistry

We have used 24 published analyses of Cornubian basalts and lamprophyres (Cosgrove 1972; Thorpe *et al.* 1986) with which to compare the larger data set of 57 Oslo basalts (Grimmer, unpubl.) - representative analyses are shown in Table 1.

Alteration effects

The variable degree of alteration exhibited by the secondary mineral development in the Cornubian and Oslo basalts has to some extent modified the chemistry of these rocks. Thus comparisons between the two regions are largely made on the basis of trace element variation, using those elements least affected by alteration, notably Zr,

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Y, P, Nb, and REE. However, REE mobility during conditions of alteration and low grade metamorphism has been described by Hellman *et al.* (1979), and Humphries (1984), although total REE mobility can often be regarded as minimal during zeolite facies metamorphism. The REE data presented shows good correlation with the more immobile HFS elements when compared with unaltered basalts of similar chemistry and are considered to reflect primary abundances. The other more mobile LIL elements show marked non-magmatic variation and have been disturbed by post-consolidation alteration.

Major chemical distinctions

Comparisons based on major elements between the Cornubian and Oslo basalts show several distinctive differences. The Cornubian basalts have overall higher levels of SiO_2, Al_2O_3 and total alkalis, and lower MgO and Fe_2O_3The rocks from both regions are alkaline in nature and have K_2O /Na_2O ratios ranging from 0.43-1.34 for the Oslo basalts and 0.27-1.61 for the Cornubian basalts. In terms of trace element abundance the Cornubian basalts have overall higher concentrations of the transition elements Cr and Ni, whereas the Oslo basalts are significantly enriched in the incompatible elements (IEs), notably Zr, Nb, Ce and P. In addition, although both groups have marked LZHREE enrichment relative to MORB, the Oslo basalts have higher overall concentrations of REEs (up to 286 ppm. against 129) ppm. for Cornubia) and enrichment of the LREEs up to 300 times chondrite. The other main distinction between

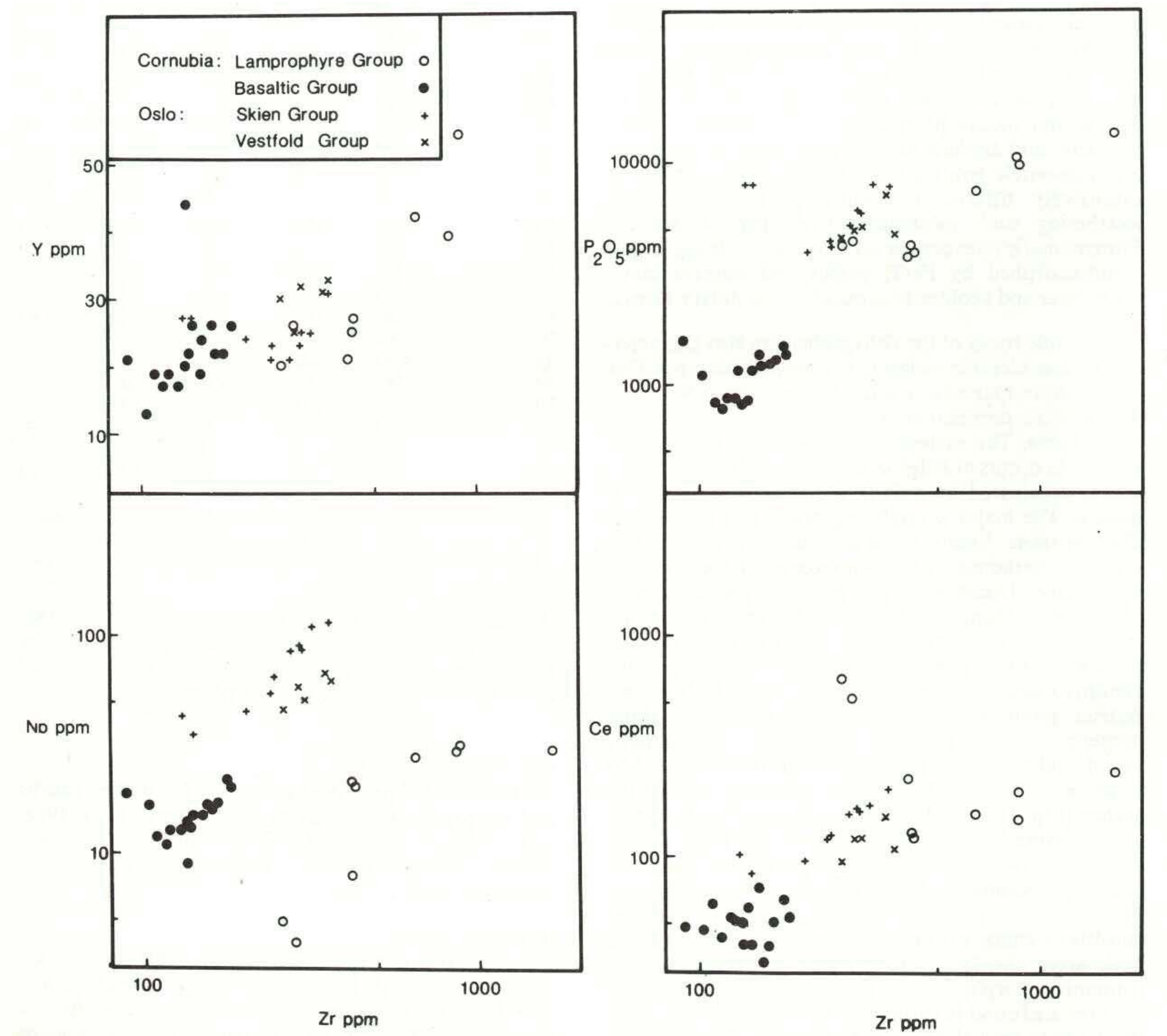

Figure 2 Incompatible element variation diagrams, Y, Nb, P and Ce vs. Zr, for Cornubian and Oslo basaltic rocks

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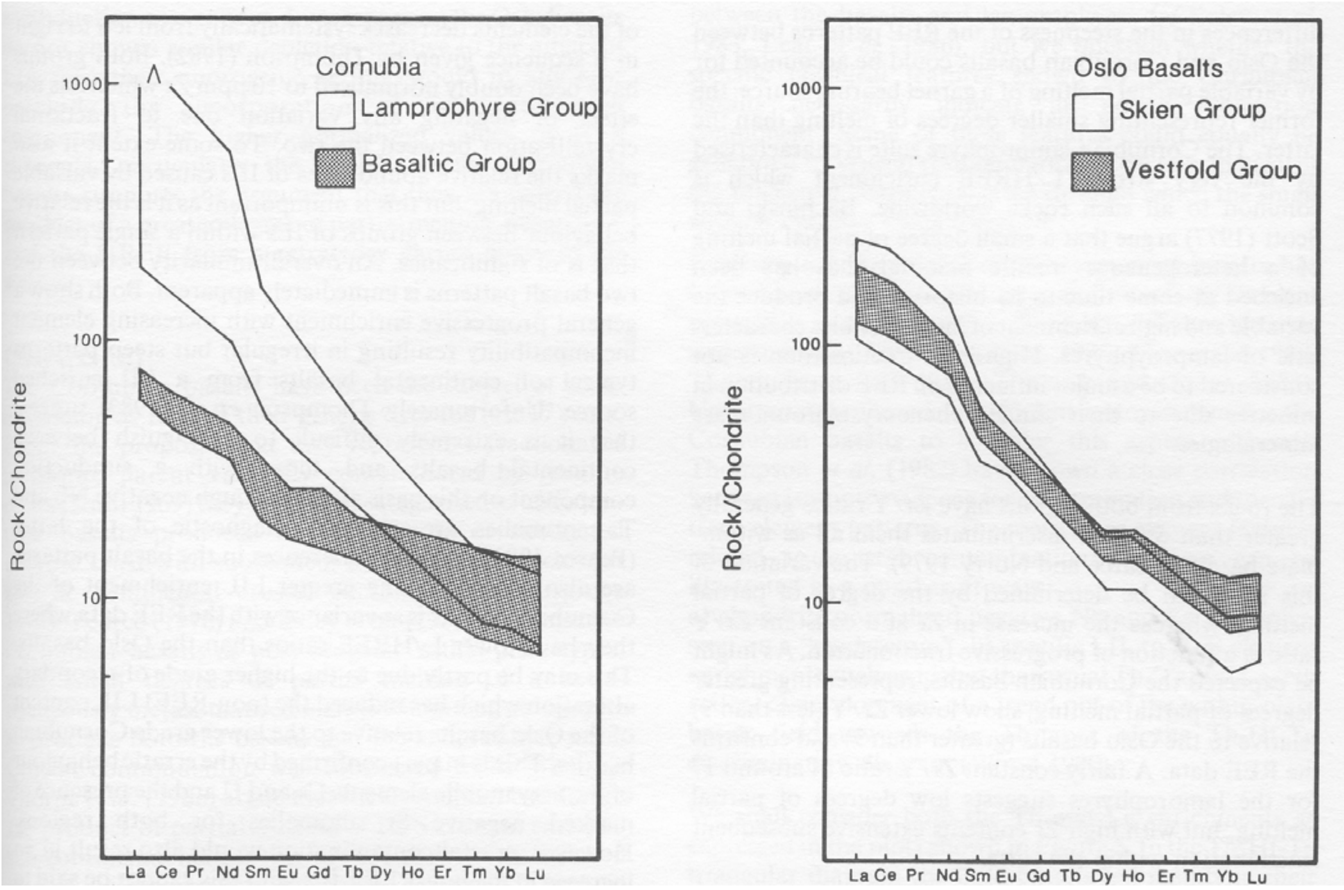

Figure 3. Chondrite-normalized REE patterns for Oslo basalts and Cornubian basalts and lamprophyres. Normalization factors from Wedepohl 1969

the two regions is the relative behaviour of the IEs. All lithological groups have highly variable LIL/HFS ratios, although the Cornubian basalts are relatively depleted in Hf, Ta and Nb. Absolute concentrations of the more incompatible trace elements, notably Rb and Th are similar.

Comparisons of petrogenetic features

Intra-suite magmatic relationships and inter-group differences due to magmatic evolution can be demonstrated by the use of the more immobile IEs. As is seen in Figure 2, volcanics from the two regions can be readily distinguished on a variety of IE plots. Furthermore, distinct comagmatic suites are recognisable for the Cornubian basalts and lamprophyres, and similarly for the Oslo basalts from Skien and Vestfold, particularly on the Ce vs. Zr and Nb vs. Zr diagrams (Fig. 2). The less evolved nature (at variance with the major element data) of the Cornubian basalts is readily apparent. The overall lower concentration of IEs in these lavas when compared with the Oslo basalts indicates a lower degree of magmatic evolution. This considerable difference in the relative "primitiveness" between the Cornubian and Oslo basalts must therefore be borne in mind when comparing the groups further. The separation from and different trend of the Cornubian basalts to the lamprophyres suggests no direct evolutionary relationship with this suite.

Variation within the various comagmatic suites from the two regions are largely related to low-pressure fractionation; many of the basalts containing observable phenocryst phases. In addition to these high-level effects differences between the Cornubian and Oslo volcanics may be the result of variable degrees of partial melting as well as source heterogeneity. For example, the strong L/HREE enrichment is a characteristic of all groups (Fig. 3) and is typical of alkali basalts derived from an enriched within-plate setting. For partial melting to produce such enrichment of the LREE in parental melts we must invoke a mantle source mineralogy with a significant proportion of minerals with relatively high distribution coefficients (Ds) for the HREE and low Ds for the LREE. Garnet, and to a lesser extent ortho-pyroxene, exhibit this characteristic with Ds ranging from 0.05-35 and 0.02-0.34 respectively in basaltic rocks (Henderson 1982). Different degrees of partial melting of such a garnet-bearing source will also fractionate the REE, with increased partial melting not only releasing a higher proportion of HREEs into the melt but also diluting the total REE content. Thus,

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differences in the steepness of the REE patterns between the Oslo and Cornubian basalts could be accounted for by variable partial melting of a garnet bearing source; the former representing smaller degrees of melting than the latter. The Cornubian lamprophyre suite is characterised by the very strong L/HREE enrichment which is common to all such rocks worldwide. Bachinski and Scott (1977) argue that a small degree of partial melting of a heterogeneous mantle material that has been enriched at some time in its history could produce the variable and high enrichment of incompatibles characteristic of lamprophyres. High-level fractionation is not considered to be a major influence on REE distribution in minettes due to their similar phenocryst/groundmass mineralogies.

The rocks from both regions have Zr/Y ratios generally greater than 5 which discriminates them all as within-plate basalts (Pearce and Norry 1979). The variation of this ratio can be determined by the degree of partial melting, whereas the increase in Zr at a constant Zr/Y ratio is a function of progressive fractionation. As might be expected the Cornubian basalts, representing greater degrees of partial melting, show lower Zr/Y (less than 9) relative to the Oslo basalts (greater than 9) and confirms the REE data. A fairly constant Zr/Y ratio of around 17 for the lamprophyres suggests low degrees of partial melting, but with high Zr contents extensive subsequent fractionation within the suite.

Figure 4 shows normalised trace element distribution patterns ("spidergrams") for basaltic rocks from the two regions. In these diagrams the degree of incompatibility the elements decreases systematically from left to right in a sequence given by Thompson (1982). Both groups have been doubly normalised to 10 ppm Zr which has the effect of negating any variation due to fractional crystallisation between the two. To some extent it also masks the relative abundances of IEs caused by variable partial melting, but this is unimportant as it is the relative behaviour between groups of IEs within a single pattern that is of significance. An overall similarity between the two basalt patterns is immediately apparent. Both show a general progressive enrichment with increasing element incompatibility resulting in irregular but steep patterns typical of continental basalts from a LIL-enriched source. Unfortunately Thompson *et al.* (1983) suggest that it is extremely difficult to distinguish between continental basalts and those with a subduction component on this basis alone although negative Nb and Ta anomalies are generally diagnostic of the latter (Pearce 1983)i Distinct differences in the basalt patterns are also exhibited. The greater LIL enrichment of the Cornubian basalts is at variance with the REE data where they have lower L/HREE ratios than the Oslo basalts. This may be partly due to the higher grade of secondary alteration which has reduced the (non-REE) LIL content of the Oslo basalts relative to the lower grade Cornubian basalts. This is in part confirmed by the erratic behaviour of the very mobile elements Cs and U and the presence of marked negative Sr anomalies for both regions. However, crustal contamination would also result in an increase in the non-REE LILs, and this cannot be said to be diagnostic (Pearce 1983). It is important to note that the LIL enrichment of the Cornubian basalts is coupled with a small negative Nb anomaly, often taken to indicate

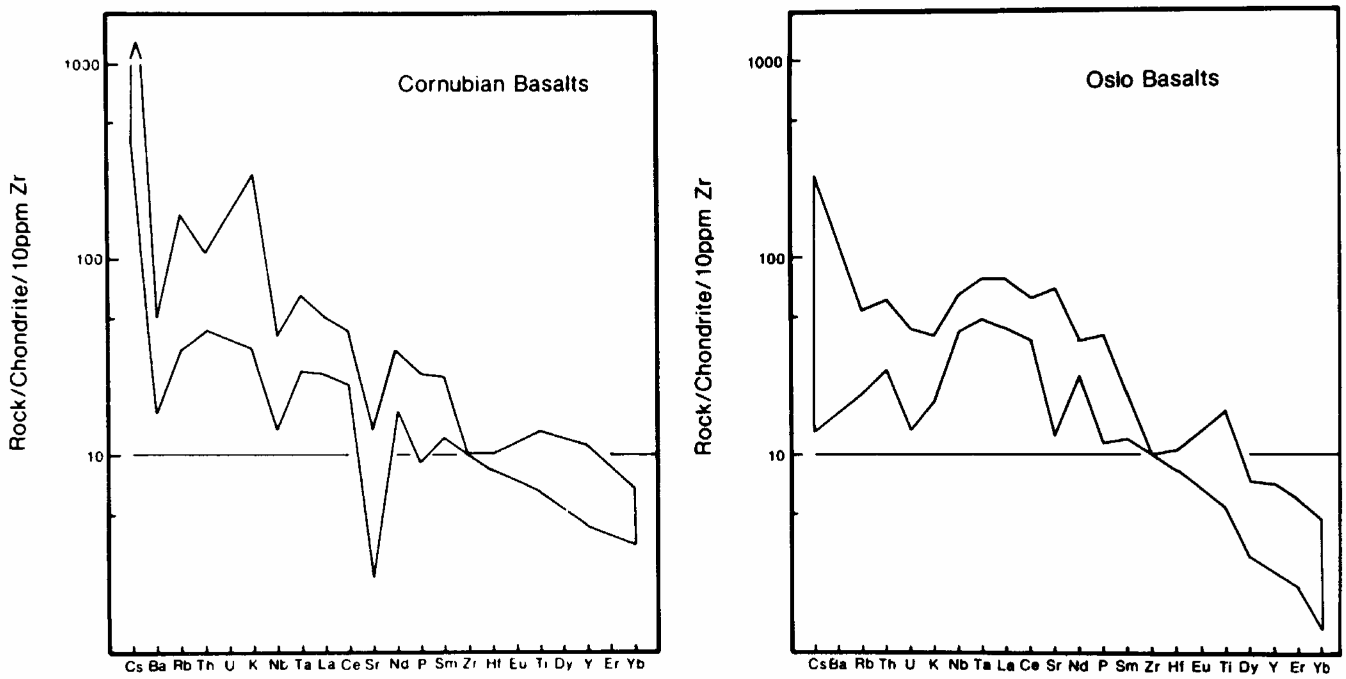

Figure 4. Trace element abundance diagrams normalized to chondrite and 10 ppm Zr for Oslo and Cornubian basalts. U not plotted for the Cornubian basalts, Ba not plotted for the Oslo basalts. Normalization factors from Thompson 1982.

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a subduction component. In comparison the Oslo basalts do not show a similar depletion relative to the adjacent LIL and their anorogenic setting would in any case preclude the incorporation of any subduction component. The higher normalized values of HFS elements, particularly the HREE, in the Cornubian basalts supports the argument that they are indeed the product of a greater degree of partial melting, releasing Y and the HREE from a refractory garnet phase in the source.

Discussion

The Cornubian basalts have been the subject of some discussion as regards their genesis over the last 17 years. From the proposal that they represent a fractionated peridotite parent variously contaminated by granitic rocks (Knill 1969) they have been categorised as collision-type basalts produced by the melting of subducted oceanic crust with subsequent metasomatic enrichment and/or continental crustal contamination (Cosgrove 1972). Pearce (1976) and Exley *et al.* (1983) recognise the Cornubian basalts as post-orogenic within-plate types, with small degrees of partial melting of a heterogeneously metasomatised mantle containing phlogopite to produce both the basalts and the lamprophyre suite. Crustal contamination was considered to be minimal. Thorpe *et al.* (1986) argue that the Cornubian basalts are the product of partial melting of HFS element enriched sub-continental mantle that has been selectively enriched in LIL elements by previously subducted oceanic lithosphere' and continent-derived sediment. The origin of the lamprophyre suite is explained by a greater degree of this subduction-related enrichment. However, Lear *et al.* (1986) suggest various mantle sources for the Cornubian suites with little modification of the source required to derive the basalt parent.

In contrast the Oslo basalts appear to be typical examples of alkaline anorogenic rift volcanism developed on relatively thin attenuated continental crust unaffected by or associated with penecontemporaneous subduction events. The Cornubian situation appears more complex as some plate tectonic models suggest the presence of a subduction zone throughout much of the Upper Palaeozoic prior to the late Carboniferous continent-continent collision (e.g. Anderton *et al.* 1979). Also, according to Thorpe *et al.* (1986) a subduction zone component is still recognisable in the chemistry of the post-orogenic Permian lavas discussed here. This feature is not seen in the Oslo basalts which have typical continental (or within-plate) alkali basalt chemistry with similar normalized Nb and Ta values to adjacent incompatible LIL elements (Fig. 4).

Thus relative to Oslo, the Permian mantle of Cornubia could be interpreted as having been "modified" via subduction processes or at least be heterogeneous in character. That some chemical heterogeneities were present would explain the chemical variations seen between the basalts and lamprophyres *(cf* Exley *et al.* 1983; Leat *el al* 1986), but we question whether the specific chemical features shown by the Cornubian basalts necessarily indicate an added subduction component. Some of these features could equally be generated by crustal contamination, especially in view of the thickened continental crust through which the small volume Permian melts travelled.

Evidence for the crustal contamination of basalts has been based on isotopic data and involving ancient crystalline basement with a distinctly different isotopic signature (e.g. Armstrong *et al.* 1977, Thompson 1982, Thompson *et al.* 1982, Thirwall and Jones 1983). Unfortunately there is insufficient isotopic data for the Cornubian basalts to consider this aspect, although Thompson *et al.* (1982) have shown a close correlation between isotopic evidence for contamination and specific trace element patterns. The main trace element features related to crustal contamination processes can be illustrated in a number of ways:
a) chondrite normalised negative Nb and Ta anomalies seen on a "spidergram", b) specific LIL to HFS element enrichments such as higher than usual Th/Ta and La/Nb ratios, c) development of a trend out of the within-plate basalt field towards the Th apex of the Th-Hf-Ta discrimination diagram (Wood 1980).

We shall consider both the Th and La enrichments as expressed in the plots shown in Figure 5. In the Th-Hf-Ta triangular diagram the Oslo basalts fall mainly in their "correct" field, that of the within-plate basalts, whereas the Cornubian basalts trend towards the Th apex and approach the calc-alkali arc field. As post-orogenic continental alkali basalts the Cornubian lavas might have been expected to fall in the same field as the Oslo data, so the displacement towards Th enrichment (or an arc environment) can be interpreted as either the presence of a subduction component in the lavas or crustal contamination. A number of continental (tholeiitic) basalts that have suffered contamination are also displaced from their "expected" within-plate discrimination polygon in this diagram, towards the Th apex. For example, the Grande Ronde Formation of the Columbia River basalts falls in approximately the same part of the diagram (Prestvik and Goles 1985, Fig. 1) as the Cornubian basalts, their position being interpreted as a possible consequence of crustal contamination.

In a similar manner the Oslo and Cornubian basalts are separated by their La/Nb ratios (Fig. 5) with the former showing relatively uniform ratios (0.73-1.09) or a horizontal trend typical of the majority of alkali basalt suites. The Cornubian basalts not only plot outside the field of continental alkali basalts (as in the Th-Hf-Ta diagram), but have highly variable La/Nb ratios (0.78-2.66) that are atypical for a single comagmatic suite. As illustrated by Floyd *et al.* (in press) some subduction-related basalts can also have enhanced La/Nb ratios (greater than l), although by comparison the relatively

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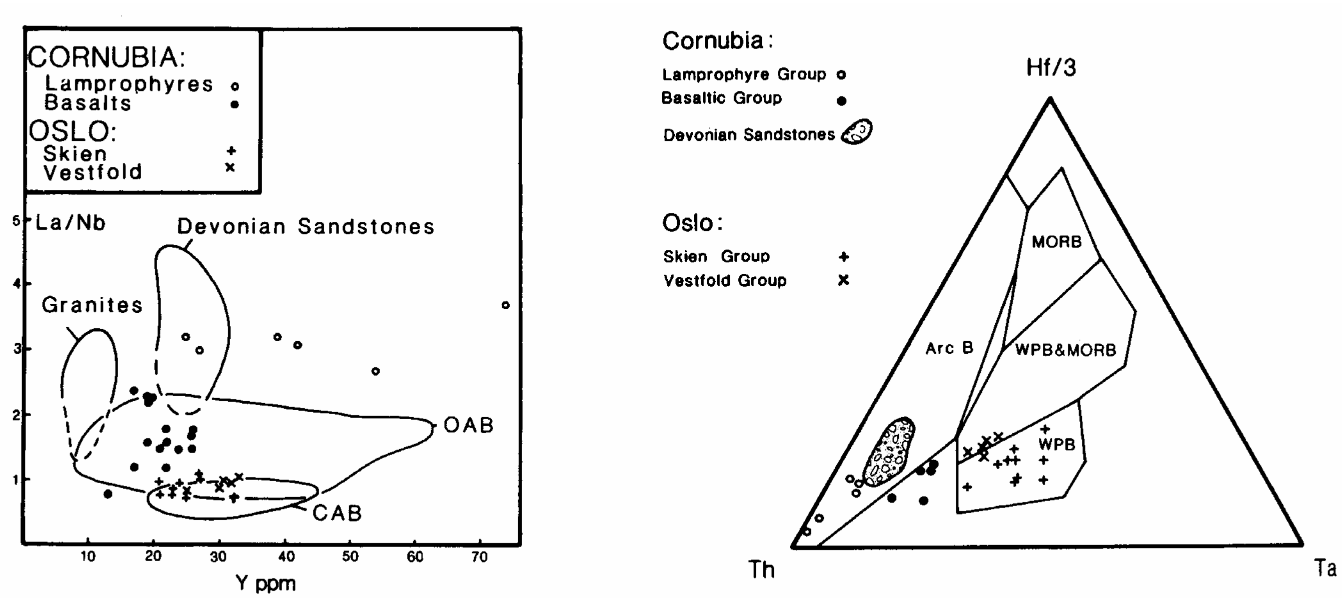

Figure 5. La/Nb vs. Y variation diagram and Th-Hf Ta discrimination diagram (Wood 1980). Compositional fields for two possible crustal contaminants from Floyd and Leveridge, 1986 (Devonian sandstones) and C.S. Exley, pers. comm. (S.W. granites). Discriminant fields for arc basalts (Arc B), mid ocean ridge basalts (MORB) and within plate basalts (WPB). Compositional fields for oceanic alkali basalts (OAB) and continental alkali basalts (CAB) from the literature.

low values seen here are more characteristic of back-arc basin environments (La/Nb 1-5) than the arc itself (La/Nb greater than 10).

Because the effects of a subduction component and crustal contamination produce similar trace element patterns it is difficult to clearly differentiate between them. In Cornubia the Permian alkali basalts are post-orogenic and continental in aspect and are not related to a contemporaneous subduction zone. Any subduction component must therefore have derived from a previous event and inherited by the Cornubian mantle. As it is by no means clear that an Upper Palaeozoic subduction zone persisted under this region, we consider that crustal contamination is an equally probable explanation for the basalt chemistry. As seen in Figure 5, local Devonian sandstones of predominantly acidic composition (Floyd and Leveridge, in press) could represent crustal contaminants of the right composition whose admixture with typical alkali basalts would produce the "intermediate" chemical features displayed by the Cornubian basalts.

Acknowledgements S. Grimmer wishes to acknowledge the receipt of a University of Keele, Department of Geology research studentship over the period of this work.

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Read at the Annual Conference of the Ussher Society, January 1986

The Bristol Channel Graben: organic geochemical limits on subsidence and speculation on the origin of inversion

CHRIS CORNFORD

Cornford, C. 1986. The Bristol Channel Graben: organic geochemical limits on subsidence and speculation on the origin of inversion. *Proceedings of the Ussher Society, 6,* 360-367.

In a study of the organic geochemistry of the Lower Lias coastal outcrops in Dorset, North Somerset and South Glamorgan a progressive maturity increase has been observed from south (Dorset) to north (Glamorgan). This trend can be interpreted in terms of higher geothermal gradients, deeper burial or a combination of both, in an asymmetric, down to the north, inner Bristol Channel graben. On regional grounds, maximum burial probably occurred by the Aptian. These findings have implications for oil exploration in the Bristol Channel and onshore as well as the structural development of the area.

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Introduction

Maturity measurements on organic components of sediments can be used to place limits on geological reconstructions (eg. Bartenstein and Teichmuller, 1971; Robert, 1985). This paper explores some of the regional implications of an organic geochemical study of limestone, marls and shales of the Lias of Dorset, North Somerset and South Wales. The detailed geochemical results are reported elsewhere together with some additional analyses (Cornford, 1972; Cornford, Forbes and Douglas, in preparation). A discussion of the tectonic framework of the area is given by Whittaker (1975) and Kamerling (1979). The interested reader is referred to the book by Tissot and Welte (1978, 1984) for further discussion of organic geochemistry.

In this study four organic maturity parameters were measured: the even carbon chain length preference of the monocarboxylic fatty acids in the range C_{25} to C_{32} termed CPIA (Kvenvolden, 1967); the odd carbon chain length preference of the n- alkanes in the range C16 to C35 termed CPIH (Scalan and Smith, 1970); the ratio of the concentrations of the isoprenoid hydrocarbons pristane and phytane termed the prist/phyt ratio. (Brooks *et al.,* 1969) and vitrinite reflectance values (Bostick 1979). An approximate equivalence of some of these maturity parameters has been discussed by Heroux *et al.* (1979). The organic geochemical methods and basic results are described elsewhere (Cornford, 1972; Forbes, 1984; Cornford, Forbes and Douglas, in preparation).

Limestones and shales were taken from the Lower Lias of Pinhay Bay, Lyme Regis, Dorset beds (H_1 to H_5 and H_{70} to H_{72} of Lang, 1924); Quantocks Head and Doniford Cliff, North West Somerset (Palmer, 1972 division A or B and C); and Lavernock Point, Glamorgan (Beds 55, 54 and 39 of Trueman, 1922). Locations are shown in Figure 1.

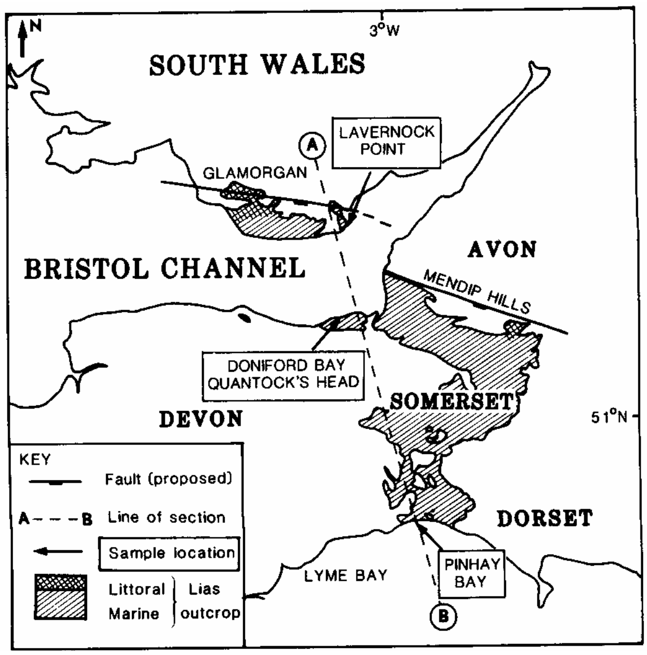

Figure 1. Sketch map of the Bristol Channel area showing the sampled locations farrowed) and the line (A - B) of the cross-sections shown in subsequent figures. The sketch map also shows a speculative north-eastern faulted boundary to a Bristol Channel/Central Somerset basin, associated with the littoral facies of the basal Lias.

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Discussion

The maturity related geochemical data are shown as a north-south cross section in Figure 2 and listed in Table 1.

The data can be interpreted in the following way. Acid and hydrocarbon CPI values, initially high in living plant waxes and recent sediments, decrease with increasing burial and hence maturity. Pristane/phytane ratios initially increase with maturity but are also affected by the provenance of the organic matter and the environment of deposition (Didyk *et aL,* 1978), the ratios being lowest in reducing environments. In Figure 2, two trends are plotted for the pristane/phytane ratios, one for sediments deposited under reducing conditions (typified by finely laminated sediments with dominantly algal and amorphous kerogens) and the other for more oxidised and bioturbated sediments showing a fairly homogenous structure and containing dominantly vitrinitic kerogens. Carbonates and shales are found in both categories.

Table 1. Some maturity parameters for the Lower Lias of Dorset, North Somerset and Glamorgan coasts.

	Vitrinite Reflect %	Carbon Index Fatty acid	Preference Hydro-carbon	Pristane/ Phytane
DORSET [2]				
PA limestone	0.35	3.2	2.1	0.8
PB marl	-	2.8	2.7	0.7
PC shale	0.37	2.8	1.7	0.4
PD marl	0.35	3.2	1.5	0.9
PE limestone	-	3.1	2.5	0.6
H_{1-5} mixed lith.	(0.28)	-	-	-
H_3 shale	0.35	-	-	-
H_{71}m marl	0.37	-	-	-
Jet' hand picked	(0.21)	-	-	-
N. SOMERSET [3]				
D_1 limestone	0.44	2.5	2.3	1.8
D_2 shale	0.44	2.5	2.1	1.2
D_3 l'stone nodule	0.48	2.7	2.2	1.7
Q_1 shale	0.46	-	0.9	0.7
Q_2 limestone	0.49	-	-	-
SOUTH WALES[4]				
L_{54} shale	0.51	2.4	1.4	2.7
L_{55} limestone	0.51	2	1.6	2.9
L_{39}S shale	(0.42)	-	-	-
L_{39}l limestone	(0.39)	1.6	1.3	0.9

1 For details of locations and analyses see Cornford, Forbes, and Douglas, in prep. () Bracketed values are not used for maturity estimations as discussed in Cornford, Forbes and Douglas, in prep.

2 PA-PE are from Lang's (1924) bed $^{H}71^{:}$ H71^{m} was collected from the same bed at a different location and date. Other Dorset (Pinhay Bay) beds labelled according to Lang's notation.

3 D, - D_3 from Doniford Bay and Q $_{1,\,Q2}$ from Quantocks Head, North Somerset coast.

4 South Wales samples all come from Lavernock Point, with bed designation after Trueman's 1922 sections.

Vitrinite reflectance determinations were made: details of the measurement are given in Cornford, Forbes and Douglas (in preparation). In all samples, particles showing a wide range of reflectance were measured, a characteristic of the Lias of North-West Europe (Hagemann, 1974; P. Flekken, pers. comm.). Because of the spread of the reflectance histograms, no reliable mean values could be readily obtained. Reworking of Carboniferous material into Lias has been reported by Wobber (1966) although no multi-macerites characteristic of reworked coal (Hagemann, 1975) were seen in this study. The average values from a number of samples at each of the three sites were 0.36%R for Dorset, 0.46%R for North Somerset and 0.51%R for Glamorgan. These results confirm the geochemical trend from south to north (Fig. 2), but, because of the spread of vitrinite reflectance values in any one sample, do not alone unequivocably define the south-north gradient of increasing maturity.

All the maturity data, if related to coal and other sediment studies, are in agreement with a vitrinite reflectance level of about 0.3% for Dorset, 0.4% for North Somerset and 0.5% for Glamorgan. As shown in Figure 3, there is little variation between the measured and concensus values in South Wales, while the Somerset and Dorset reflectance values indicate a higher degree of

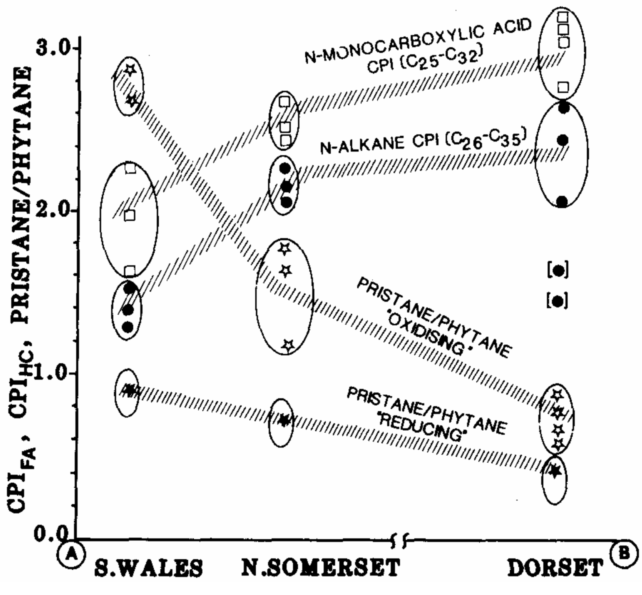

Figure 2. Variation of maturity controlled geochemical parameters from Dorset to South Wales (see Figure 1 for line of section). The carbon preference indices (CPIs) of the n-monocarboxylic acids in the range C_{25} - C_{32} (squares) and the n-alkanes in the range C_{26} - C_{35} (solid circles) decrease with maturity towards the north. Pristane/phytane ratios (crosses) of both the laminated lithologies containing dominantly algal/amorphous kerogens (reducing depositional environment) and the homogenous or bioturbated lithologies containing dominantly terrigenous land plant kerogens (more oxidising depositional environment), increase with maturity to the north. Square bracketed vanes of n-alkane CPIs from Dorset samples are unreliable due to low amounts of alkanes.

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maturity than the extract parameters. Given the broad spread of VR measurement obtained from any one sample (from low reflecting "Jet" to semi-fusinite), the maturity parameters taken from the extracts are taken as definitive. The discussion in this paper is therefore based on the concensus maturity levels.

Two Models

Using these maturity data a rough calculation can be made of the relative burial/temperature histories of the three outcrops. Two extreme models can be considered:

1) Uniform geothermal gradients and differential burial

At Pinhay Bay 100m of incomplete Lias section is overlain by about 140m of Upper Cretaceous (Edmonds *et aL.,* 1975). Further east in Dorset a more or less complete section of Jurassic and Cretaceous sediments is found (Melville and Freshney, 1982). Assuming a similar complete Jurassic section overlay the Lias of Pinhay Bay, and allowing for 100m of Lower Cretaceous burial prior to uplift and erosion, a maximum burial of some 1400m is indicated. These calculations assume Lower Cretaceous uplift and erosion to allow Gault (Albian) deposition directly upon the Lias as seen today.

As Tertiary cover was probably minimal (Edmonds *et al.,* 1975), maximum burial experienced by the Pinhay Bay Lias occurred in the pre-Albian early Cretaceous. Travelling east into the Wessex basin, maximum burial would have occurred progressively later.

A similar calculation in the Bristol Channel gives a maximum burial of about 3420m for the East Bristol Channel by the Aptian (Kamerling, 1979). Cope (1984) has however put a case for the minimal Lower Cretaceous burial in the inner Bristol Channel giving about 2250m burial for the base Lias by mid Cretaceous times.

A straight line relationship is normally obtained if vitrinite reflectance is plotted on a logarithmic scale against maximum burial depth on a linear scale (Dow, 1977). Using the normally observed surface vitrinite reflectance of 0.2%R, the 0.3%R at 1400m maximum burial in Dorset extrapolates to about 2400m maximum burial for 0.4%R in North Somerset and 3200m burial for 0.5%R in South Wales (Fig. 3). These depth values should be taken as maxima but seem to confirm the regional estimates for the East Bristol Channel basin which include a fairly thick Lower Cretaceous section (Kamerling, 1979).

A further upper limit on Mesozoic burial is indicated to the north of the Bristol Channel by the presence of the Carboniferous high volatile bituminous coals (ca 33% vol matter and about 1.0%R) on the Eastern rim of the South Wales Coalfield (Gill *et al.,* 1979). The mature pattern of the Carboniferous, presumably attained during the Variscan burial has thus not been overprinted during any subsequent Mesozoic burial.

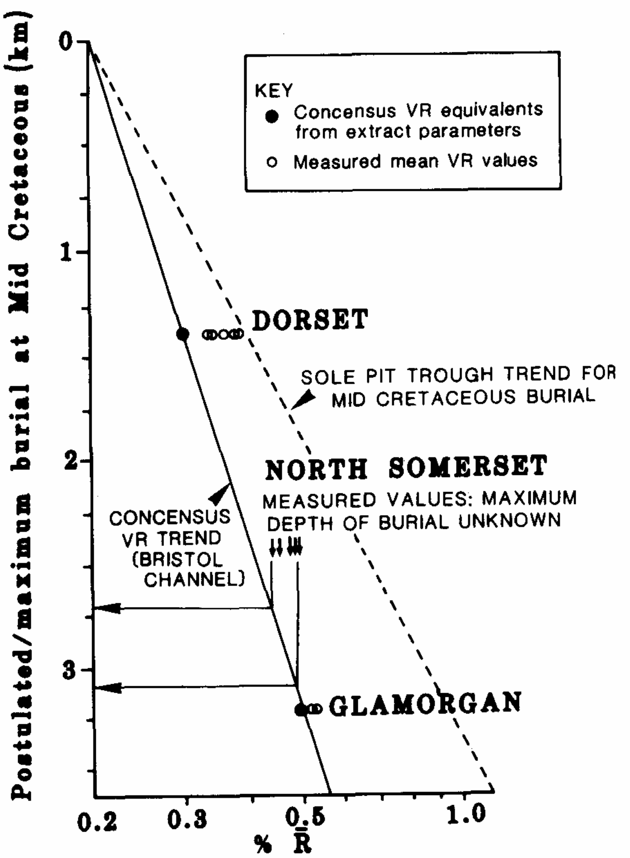

Figure 3. Measured vitrinite reflectance values (open circles) for the Dorset and South Wales basal Lias plotted against their respective Lower Cretaceous (pre-Aptian) estimated maximum depths of burial. The solid circles represent the concensus values of VR equivalents derived from the extract data. A conventional surface intercept of 0.2%R is utilised. The reflectance value of 0.4%R for North Somerset indicates pre-Aptian paleo-burial of about 2.4km. Note that the reflectance values are plotted on a logarithmic scale.

2) Uniform burial and changing geothermal gradient

An alternative explanation of the maturity related data is the combination of more uniform burial and variable geothermal gradients. Using the approach of Hood *et al* (1975), an approximate equivalence of vitrinite reflectance and temperature can be made for rocks buried near their maximum depth - or more correctly within 15°C of their maximum temperature - for say 10m.y. (i.e. an effective heating time of 10m.y.). Using an effective heating time of 10 million years for the Early Cretaceous maximum burial event the following temperatures can be obtained from the concensus vitrinite reflectance values (Bostick, 1979):

0.3%R = 35°C;
0.4%R = 55°C;
0.5%R = 80°C

Higher temperatures would be needed to reach a given vitrinite reflectance level if the time of maximum burial

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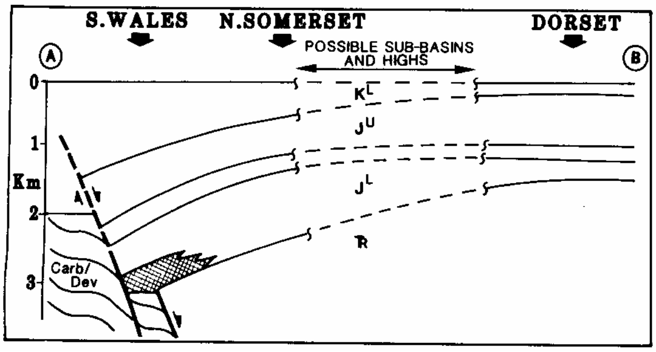

Figure 4. Speculative sketch section from Dorset to South Wales suggesting pre-Aptian burial consistent with the geochemical maturity indicators from the basal Lias.

(the effective heating time) is shorter. Assuming 3200m maximum burial for both the Bristol Channel sites and a 15°C surface temperature during the Early Cretaceous, geothermal gradients of 13°C/km and 20°C/km would be indicated for the Somerset and Glamorgan sites respectively. A geothermal gradient of 14°C/km is indicated for the Dorset section during the Early Cretaceous.

These gradients are somewhat lower than expected, by analogy with present day continental margins (eg. North Sea, Cornford, 1984). Assuming less Lower Cretaceous sedimentation (Cope, 1984) and hence maximum burial of 2250m in the Bristol Channel gives gradients of 18°C/km and 29°C/km for Somerset and Glamorgan respectively. These values, though still low, are less extreme by present day standards.

This analysis hence does not favour a model where maturity changes are solely a function of different geothermal gradients at the three locations.

Mesozoic -Tertiary grabens for example in the North Sea, and Upper Rhine areas are accompanied by a) abnormally thick sequences (Zeigler, 1982) and b) higher geothermal gradients (Sclater and Christie, 1980). The geothermal gradients operating during the Mid Cretaceous maximum burial event can only be guessed at. The maturity of the Carboniferous (Westphalian) strata of the Sole Pit Trough (Barnard and Cooper, 1983) attained during Cretaceous burial (Glennie and Boegner, 1981) suggest reflectance values of 2% with burial of about 5km. This maturity gradient is also marked in Figure 3. Like the Bristol Channel area, the Sole Pit Trough also suffered inversion in the mid Cretaceous.

Thus model 1 based on differential burial, or the somewhat less plausible model 2 based on variable paleo-geothermal gradients, or more likely a combination of both, point to the development of a Mesozoic graben in the Bristol Channel area.

Time/Temperature Modelling

This technique (Waples, 1980) can be used to combine the elements of the two models discussed above. In essence the modelling follows the burial of any given stratum as it is exposed to progressively higher temperatures under the influence of geothermal gradients that may change with geological time. The maturation process is modelled as approximating to a first order chemical reaction, where the effects of temperature and the time of exposure to that temperature, are combined to yield TTI (Time-Temperature Integral) units. These TTI units have been empirically related to vitrinite reflectance and other measures of maturity (Waples, 1980; Issler, 1984).

Using the sequence of events listed in Table 2, the time/temperature history of the Liassic strata can be reconstructed (Fig. 5). If a graben model is used, then crustal thinning, listric faulting and rapid subsidence are evidenced by the thick (-3000m) section of Liassic to pre-Aptian sediment in the inner Bristol Channel. Geothermal gradients at this time during the Jurassic would be predicted to be initially high due to the thinner crust (Sclater and Christie, 1980). Kamerling (1979) has suggested from seismic, lithostratigraphic and regional evidence that maximum maturity was reached by about Aptian time. The Cretaceous and Tertiary thermal subsidence phase would be expected to give rise to slower sedimentation rates together with a falling geothermal gradient. No additional maturity need necessarily accrue under these conditions.

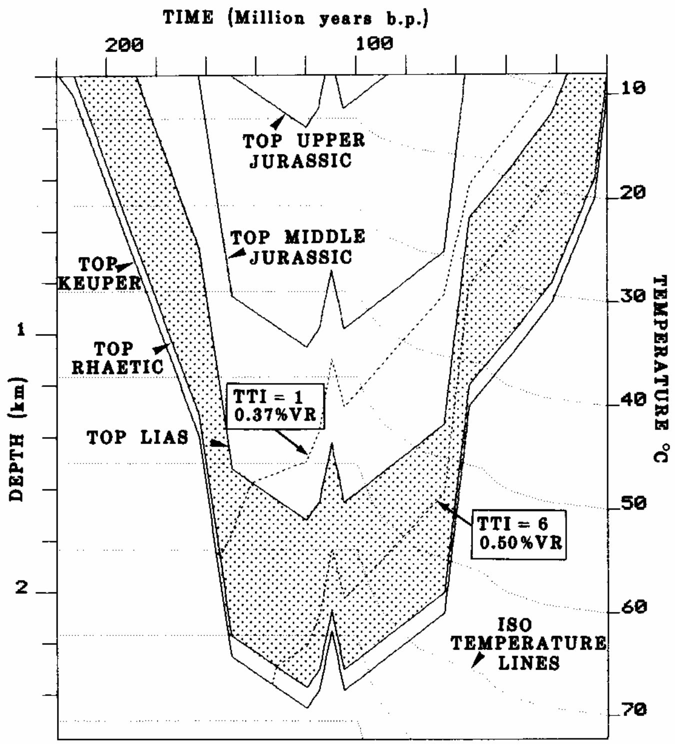

Figure 5. Subsidence history and time/temperature maturity modelling of a hypothetical section in the mid Bristol Channel. The input details are discussed in the text and in Table 2. The Lias is modelled as reaching a calculated reflectance equivalent to 0.5% by the early Cretaceous, with a present day value of 0.54% after uplift.

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Time-temperature modelling of the Bristol Channel data (Fig. 5) can be undertaken using thicknesses discussed in Cope (1984) and Kamerling (1979) together with the following assumptions:

- A thickness of 80m of Lower Cretaceous as suggested by Cope (1984) but thought an underestimate by Kamerling (1979).
- A Present day geothermal gradient and surface temperature of 20°C/km and 8°C respectively.
- An Eocene surface temperature of 12°C based on the paleoclimate and paleolatitude.
- A Jurassic geothermal gradient of 30°C/km and a surface temperature of 15°C based on the crustal thinning model and the paleoclimate and paleo-latitude.
- Uplift to allow total erosion of the Upper Cretaceous and part of the Jurassic by Eocene times (see below).

This modelling yields a predicted present day Time - Temperature Index (TTI) values (Waples, 1980) of 10 for the Lower Lias, equating with a vitrinite reflectance value of 0.54 %R using the VR/TTI equivalence of Issler (1984).

This is not a unique combination of values: many other permutations and combinations of geothermal gradients and thicknesses will give a similar result. It does not however suggest a considerable thickness of Lower Cretaceous strata, if geothermal gradients are to be kept within reasonable bounds.

Regional burial and uplift

In terms of the regional geology this pattern of higher maturity in the South Wales area could correlate with an extention of the Mesozoic South Celtic sea trough into the Bristol Channel area. Seismic results, coupled with what sparse evidence is available for the submarine outcrops in the inner Bristol Channel, indicate up to 2100m of preserved Jurassic in the synclines and inconclusive evidence for about 1100m of pre-Aptian Cretaceous deposits (Kamerling, 1979 and references therein).

It has been argued that intrusion of the Lundy granite in the Eocene indicates a thick cover by that time (Lloyd *et al.,* 1973), but the presence of reworked Jurassic palynomorphs in the Eocene Stanley Bank basin sediments (Boulter and Craig, 1979) shows that uplift and erosion down to Jurassic level had already occurred by Eocene times. Also the lack of Cretaceous flint pebbles in the Stanley Bank Basin sediments indicates total chalk erosion in the area by this time (Cope, 1984). A plausible sequence of events is detailed in Table 2.

It has been pointed out that there is no evidence for a littoral facies in the Lias of the North Somerset coast (Palmer, 1972). By analogy with the onlap of the Lias over the London platform to the east (Donovan *et al.,* 1979) an abrupt margin may not have existed. Indeed

Table 2 Proposed sequence of events affecting Liassic subsidence in the Bristol Channel area.

AGE	EVENT	EVIDENCE
213-188my	Lias deposited	Mid channel section and on flanks of Exmoor'; highest sedimentation rates in North Somerset: no evidence for shoreline'
188-144my	Mid and Upper Jurassic section deposited	Mid-channel section','
144-120my	Lwr Cretaceous deposited	Of speculative thickness','
120-100my	Minor uplifts and erosion	Presence of chalk overlying Paleozoics eg. Orleigh Court pebble beds, North Devon'.
100-65my	Chalk deposited over whole area	Remanie deposits'
65-55my	Chalk eroded	Complete stratigraphic section represented in Orleigh Court pebble beds, North Devon', no flints in Stanley Bank Basin sediments'.
55-38my	Jurassic eroded	Jurassic Palynomorphs in Stanley Bank Basin Eocene4 Radiometric dating5.
38-present	Further Jurassic erosion	Presence of basal Lias/top Trias at surface on flanks of basin 1,6

1 - Cope, 1984
2 - Whittaker, 1976, Kamerling, 1979
3 - Rogers and Simpson, 1937
4 - Boulter and Craig, 1979
5 - Lloyd et al., 1973
6 - Palmer, 1972

Whittaker (1976) shows that at least part of Exmoor accumulated normal offshore Liassic sedimentation.

In contrast, the northern margin of the graben was probably abrupt, lying to the north of the Glamorgan coast possibly on or about the zone of the "littoral" facies of the Lower Lias (Hallam, 1960). Thus, the geological and maturity data suggest that the graben was sedimentologically, structurally and/or thermally asymmetric, possibly a half graben with the deepest portion or highest gradients adjacent to the northern bounding fault along the Glamorgan margin (Fig. 4).

The position of the northern boundary should lie to the north of the Lavernock outcrop, and not to the south as indicated by Whittaker (1975). It may be coincidence that it appears to parallel the tectonic manifestations of the Variscan Front.

A possibility that flies in the face of current opinion (e.g. Cope, 1984), but little fact, is that there was no northern boundary to the Bristol Channel depocentre during the Jurassic to Cretaceous, and that a thick Mesozoic cover existed over Wales. Alridge (1986) invoked a thick Mesozoic cover over Wales as one possible explanation

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of conodont colour maturity indices in the Lower Paleozoic. Thus Figure 4 could project to the north into the very thick Jurassic sequence of the Mochras Borehole in North Wales.

A half-grabenal structure for the inner Bristol Channel basin has, however, already been suggested (Whittaker, 1975; Kamerling, 1979). This imbalance in subsidence was reversed during the Early Jurassic since it has been noted (Palmer, 1972) that the North Somerset coast section has higher sedimentation rates than either the Dorset or South Wales sections.

Timing of uplift

The outstanding question is the timing of the major phase of uplift and erosion that has exposed the once relatively deeply buried Lias. Inversion implies moving from a tensional to a compressional regime. Cretaceous and Palaeocene phases of inversion have been suggested for St. Georges Channel in the Irish Sea (Dobson *et al.,* 1982). To the south, late Alpine (Oligo-Miocene) compressional structures have been noted (eg. Colter and Havard, 1981).

The axes of the inner and outer Bristol Channel synclines and the central Bristol Channel fault zone are offset by the NW-SE trending faults paralleling the Sticklepath fault (Lloyd *et al.,* 1973). The Sticklepath fault was active at least up to the Oligocene as evidenced by the Eocene age of the Lundy granite and the Eocene-Oligocene Stanley Bank basin sediments (Boulter and Craig, 1979; Freshney *et al.,* 1982). This points to uplift, faulting and folding in the Upper Cretaceous-Palaeogene (?pre-Eocene, Evans and Thompson, 1979) rather than in the Neogene. The origin of the late Cretaceous/Palaeogene compressional forces is not clear, but it is noticeable that Cretaceous sediments are missing to the north west in Cardigan Bay - St. Georges Channel (Barr *et al.,* 1981), and are present but with disconformities in the Celtic Sea and Fastnet basins (Naylor and Shannon, 1982).

Since, to an approximation, every compressional inversion must have a conjugate extension in order to conserve volume, we can ask ourselves why did the Bristol Channel subside so rapidly during the Jurassic, and reverse its direction of motion during the Cretaceous and Tertiary? The extension of the Jurassic and Lower Cretaceous reasonably correlates with the early rifting of the Atlantic suture to the west. This dominantly southeast-northwest tension could have been reoriented into an essentially north-south direction by older Variscan lineations.

The inversion, however, should be seen in the context of mid- late Cretaceous sea floor spreading in the Atlantic to the west, the latter stages of the opening of the Bay of Biscay to the south (Soler *et al.,* 1981) and the early phases of the Tethyan Alpine margin collision to the south east. To the north and east, the North Sea graben system was undergoing extension, with the locus of extension migrating northward with time.

By the Upper Cretaceous the North Viking Graben (Hancock, 1984), and every more dramatically the More Basin (Hamar and Hjelle, 1984), was actively subsiding. It seems possible that the late Cretaceous-Paleogene compression was caused by the anti-clockwise rotating of the UK - Eire block about an axis in the area of the Southern North Sea, actively widening the Viking Graben/More Basin in the north and compressing the basins of the Bristol Channel, and Cardigan St. Georges Bay. Such a compressional phase also affected the Sole Pit Trough/Cleveland Basin, and in a basement subsidence sense the Wessex basin (Chadwick, 1985).

General Comments

These conclusions have two important corollaries, one economic, one academic. In terms of economics, the presence of the Lias shales, in part excellent petroleum source beds, approaching, in the South Wales outcrop, petroleum generation rank is important for onshore and oil offshore oexploration in the region. However, the early (pre-Aptian) attainment of maturity does not favour the presence of hydrocarbons since the survival of the reservoired hydrocarbons through the period of inversion to the present day would require an exceptionally well sealed structure.

Of more academic significance is the implication that the bioclastic limestones of the so called "littoral" facies Lias of the Sutton and Southern Down Stone in South Wales and the Downside Stone of Shepton Mallet, North Somerset may be associated with the northern faulted boundary of a half graben, possibly accumulating at a paleo-fault scarp. There is, however, no evidence for a major fault visible on the ground in the South Wales area though the lineation does appear to follow one manifestation of the Variscan Front. Outcrops are poor, and after inversion such a structure may not show up on geophysical surveys.

The present day outcrop of the "littoral" facies Lias trends approximately east-west, ie. parallel to the other major structural elements of the region (Lloyd *et al.,* 1973; Whittaker, 1975; Brooks and Al-Saadi, 1977). The alternative to a faulted northern boundary is continuous thick sedimentation over South Wales as mentioned above.

In addition, the other outcrop of the "littoral" facies Lias, the Downside Stone of the eastern Mendips, falls on a mappable extension of the major Beacon Hill fault. Here detailed field mapping indicates intra-Early Jurassic fault movement (Cornford, unpublished mapping). Intra-Early Jurassic fault movement is also evidenced by the movements on the Radstock shelf (Tutcher and Trueman, 1925) and opening of fissures in the Eastern Mendips (Copestake, 1982).

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It is possible that the major tectonic lineation noted to the south of the Mendips (Green and Welch, 1965) forms the northern boundary of the same half graben, which would therefore include the central Somerset basin (Fig. 4). Whittaker (1973) describes the central Somerset basin as a relatively symmetric steep-sided graben with active faulting in the Early Mesozoic. Tectonic inversion may be responsible for the present day symmetry. Also, Macquaker *et aL* (in press) have reported an anomalously high maturity for the Burton Row borehole in the centre of the Somerset basin, further supporting major axial burial.

Finally, the intriguing presence of the Middle Jurassic Fullers Earth deposits of the Bath area, montmorillonite clays of presumed volcanic origin, may also be indicators of nearby igenous activity associated with the deep seated faulting of an inner Bristol Channel - central Somerset half graben active at this time.

This type of organic geochemical maturity analysis is a powerful regional tool for tectonic reconstruction. Its real power lies in the presence of a grid of data points. Additional data have been generated from further Lias outcrop samples by Forbes (1984) and from outcrop and borehole samples from the Rhaetic by Macquaker *et al.* (in press). When placed together in a regional pattern this information should clarify the extent of the Bristol Channel Graben and place further limits on the timing and magnitude of the inversion episode.

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Read at the Annual Conference of the Ussher Society, January 1986

The late Triassic succession in central and eastern Somerset

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Warrington, G., Whittaker, A. and Scrivener, R.C. 1986. The late Triassic succession in central and eastern Somerset. *Proceedings of the Ussher Society, 6,* 368-374.

The late Triassic succession, from the upper part of the Mercia Mudstone Group to the basal Lias, is poorly exposed in central Somerset and is largely concealed beneath younger deposits in the eastern part of the county. Boreholes near Langport, central Somerset, proved Mercia Mudstone Group deposits (Blue Anchor Formation (c.19m), on red mudstones) succeeded by the Penarth Group (c. 13m) and basal Lias. A thinner sequence, in which red mudstones (7.9m) resting unconformably upon Carboniferous Limestone are succeeded by the Blue Anchor Formation (6.6m), Penarth Group (10m) and Lower Lias, has been proved farther east in a borehole at Bruton. Late Triassic palynomorphs have been recovered from the Blue Anchor Formation and Penarth Group near Langport.

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Introduction

Excellent exposures of the late Triassic Mercia Mudstone Group to basal Lias succession are to be seen on the west Somerset coast (Mayall 1981; Whittaker and Green 1983; Warrington and Whittaker 1984). Inland in the county, few significant exposures of these beds remain though railway cuttings and quarries formerly afforded important sections (Woodward 1893, 1905; Richardson 1911). The sequences present in central and eastern Somerset have, however, recently been proved in boreholes near Langport and Bruton respectively.

Langport

The youngest Triassic and oldest Jurassic beds around Langport, Somerton and Street in central Somerset have been neglected by geologists for many decades. Even when active quarries existed (Woodward 1893) the relationships between the worked beds and contiguous lithostratigraphical units were commonly unclear. Important sections in railway cuttings at Langport and Charlton Mackrell were described by Woodward (1905) and by Richardson (1911), but exposure in the area is poor at the present time.

In 1973 a series of cored boreholes (R1-R13; Fig. 1) was drilled for the Wessex Water Board in the High Ham area about 2 km north of Langport, where plateau-forming Lower Lias limestones are dissected by a shallow valley that cuts down through the Penarth Group to expose the Mercia Mudstone Group.

The late Triassic succession from the basal Lias pre-planorbis beds down to beds just below the Blue Anchor Formation in the Mercia Mudstone Group was encountered in the boreholes which were logged by A.W. and R.C. S. The majority were not cored from the surface and only the cored sections are represented and described here (Fig. 2). The High Ham boreholes are near the southern side of the Central Somerset Basin (Whittaker 1973); they are some 20km south-west of outcrops of the same succession near Wells (Duff *et al.* 1985) and are approximately equidistant between those on the coasts of south Devon (Richardson 1906) and west Somerset. Unless otherwise indicated, the following lithostratigraphical descriptions are based upon records from all the relevant boreholes.

Lithostratigraphy:

Lias

The basal Lias beds, cored only in R 13, are typical of that sequence in south-western Britain with dark bluish grey, fine-grained, hard limestones, with some small bivalves, in beds from 0.14 to 0.47m thick that alternate with medium to dark grey and brownish grey, slightly shalt' or fissile mudstones.

Penarth Group

This Group, comprising the Lilstock Formation overlying the Westbury Formation, was not completely cored in any one borehole; the fullest record is from R4.

The Lilstock Formation, comprising the Langport Member overlying the Cotham Member, was fully proven in R13 and partially in R4 and 8.

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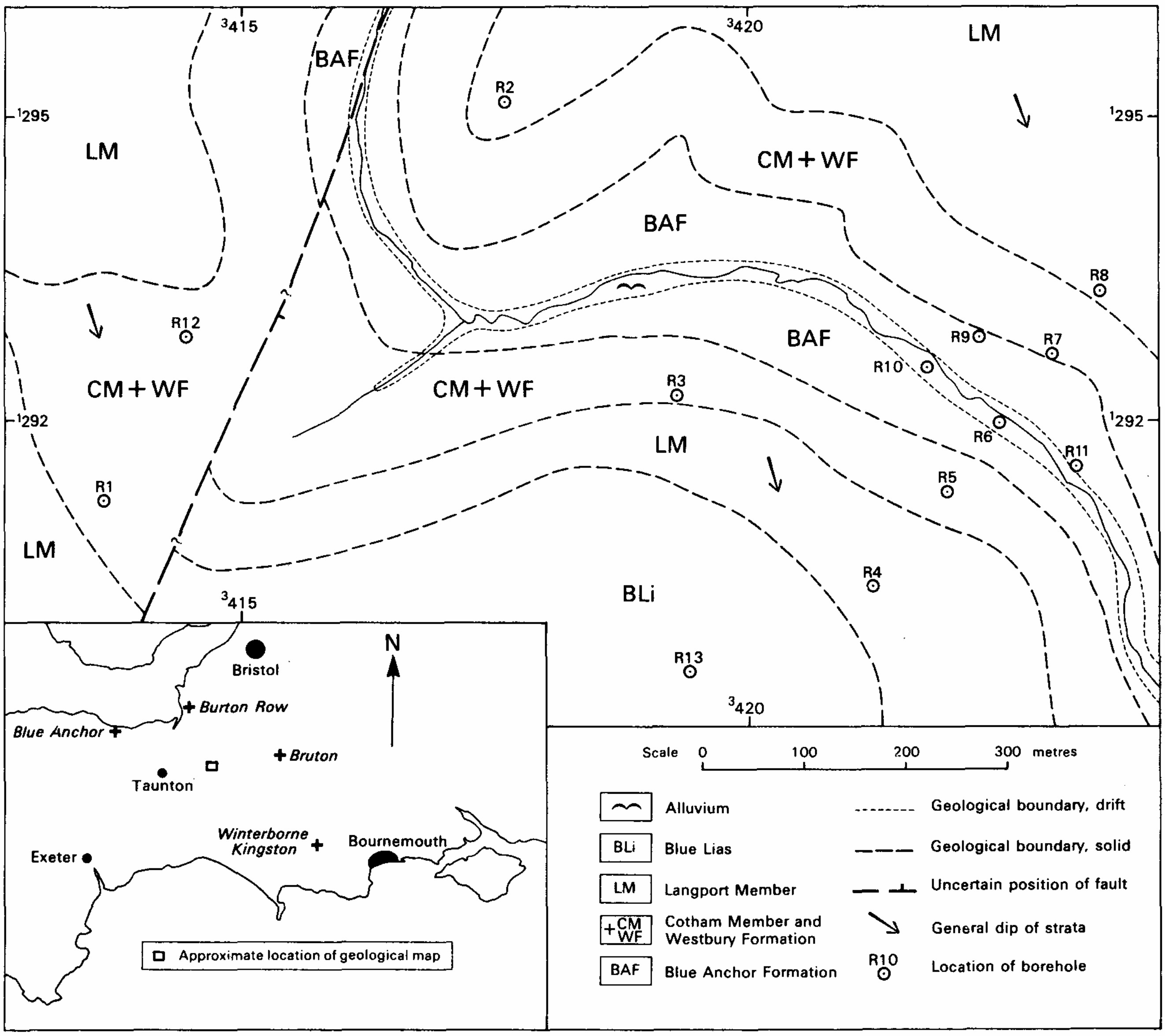

Figure 1. Geology of the High Ham area near Langport, central Somerset, and location of Wessex Water Board boreholes RI-R13 (geology after A. Whittaker). Inset map: location of the High Ham area and other sites mentioned in this account.

The Langport Member is 4.69m thick in R13 and comprises pale grey, fine-grained or porcellanous limestones in beds up to 0.61m thick that alternate with beds of pale grey calcareous mudstones up to 0.9m thick.

The Cotham Member is 1.5m thick in R4, and 1.22m in R13 where 0.9m of medium to dark greenish grey mudstone, with wisps and partings of grey silt, overlie 0.32m of medium to dark grey calcareous mudstone and siltstone; its absence from the R8 section (Fig. 2) is attributed to cambering of the superincumbent Langport Member on the valley side at that site (Fig. 1).

The Westbury Formation is 7.07m thick in R4 and 7.22m in R8. It comprises dark grey pyritous shalt' mudstones with some dark grey, fine-grained Shelly limestones, in places with fibrous calcite ("beef"); the lowest beds (0.12m) are arenaceous.

Mercia Mudstone Group

Twelve boreholes yielded sections of the Blue Anchor Formation, and lower beds were reached in four boreholes (Fig. 2).

The Blue Anchor Formation, comprising the ' Grey Marl' overlying the 'Tea Green Marl' (Whittaker and Green 1983; Warrington and Whittaker 1984), was fully proven in R8, where the following succession was recorded (thicknesses in metres):

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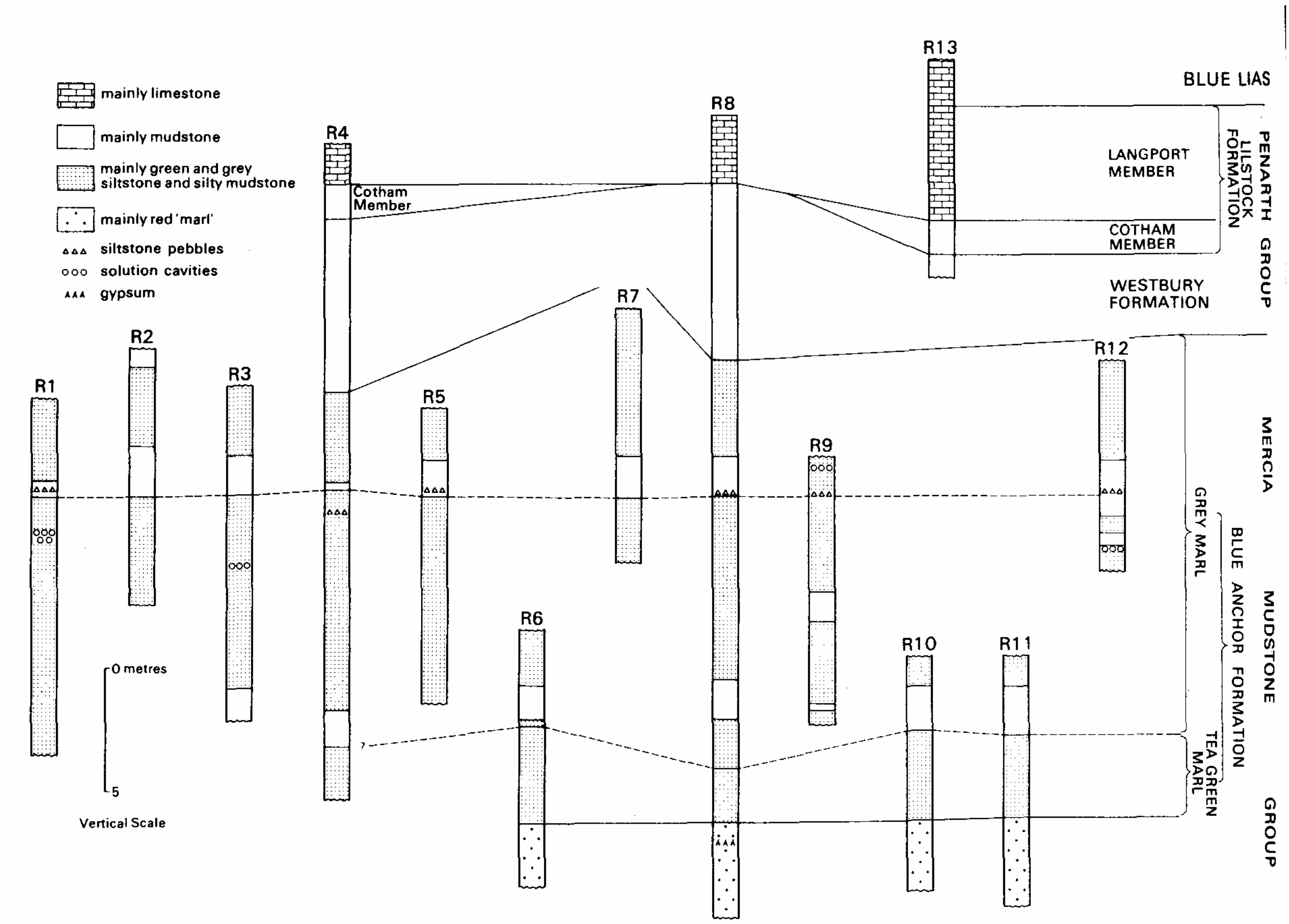

Figure 2. Successions cored in the High Ham boreholes.

Mercia Mudstone Group

BLUE ANCHOR FORMATION

Grey Marl (16.74m):

Mudstone, greenish grey and green, silty, blocky; some pale grey siltstone partings and patches of dark grey mudstone imparting a striped appearance locally 3.95

Mudstone*, very dark grey; some hard pale grey siltstones; angular pale grey siltstone pebbles in 0.12-m unit near base1.63

Siltstone, medium grey and green; some dark grey mudstones; locally fragmented 2.46

Mudstone, medium greyish green, blocky, marly; pale greenish grey siltstones to 0.1m; traces of red clayey marl 5.09

Mudstone*, very dark grey; numerous thin sub-horizontal gypsum veinlets 1.61

Siltstone, buffish grey; patches of green marl 0.61

Mudstone, greyish green, blocky, marly; some pale buffish grey siltstones; thin gypsum veinlets dipping at 45° 1.29

Mudstone, dark grey; with wisps and mottled patches of green marl 0.10

Tea Green Marl (2.2m):

Mudstone, greyish green, blocky; thin siltstones (to 0.06m) and mudstone-filled cracks 0.84

Mudstone, dark greenish grey, slightly fissile; very thin gypsum stringers in places 0.48

Mudstone, medium and dark greenish grey, marly and silty0.45

Siltstone, medium grey to greenish grey; hard and blocky in upper 0.17m, marly below 0.40

(red siltstone: *see below)*

The 'Grey Marl' sequence includes two very dark grey mudstone beds (* in the above section), each c. 1.6m thick in R8, that are recognisable in many of the other sections recorded at High Ham and may correlate with the units designated B and C in sections on the west Somerset coast (Whittaker and Green 1983; Warrington and Whittaker 1984). The upper dark grey bed incorporates angular dark grey siltstone intraclasts and the underlying sediments show disturbed bedding in Rl, 5, 8 and 12; siltstone intraclasts occur at similar levels in R4 and 9. Cavities, up to 0.01 m in diameter, that probably result from solution of evaporites occur in R1, 3, 9 and 12 and impart a honeycomb appearance to the host rock; some are lined with calcite or associated with small-scale calcite veining.

The 'Tea Green Marl' varies in thickness from 2.2m (R8) to 4.04m (R6). Bioturbation, minor bedding dislocation and local dolomitisation occur.

The Blue Anchor Formation is thinner at High Ham than at the type locality in west Somerset (Table 1) and lacks the nodular and bedded sulphates that are prominently developed there; the sequences are, in other respects, lithologically identical.

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Table 1. Thicknesses of late Triassic deposits in Somerset.

	Blue Anchor[1]	High Ham[2]	Bruton[3]
Langport Member	1	4.69	6.7
Cotham Member	1.5	1.22-1.5	1.5
Westbury Formation	14	7.07-7.22	1.8
Blue Anchor Formation	36.54	18.94	6.6
Red beds in Mercia Mudstone Group	> 67.0	> 3.90	7.9

Penarth Group thicknesses from Richardson (1911); Blue Anchor Formation from Warrington and Whittaker (1984); red Mercia Mudstone Group (minimum) thickness from thickest sequence exposed near Blue Anchor (St. Audrie's Bay; Whittaker and Green 1983).

[2] this account.

[3] from Holloway and Chadwick (1984).

The Mercia Mudstone Group succession below the Blue Anchor Formation was penetrated for a few metres only in R6, 8, 10 and 11; the following beds were recorded in R8 (thicknesses in metres):

(Blue Anchor Formation)

Siltstone, hard, purplish red and purplish grey, blocky; scattered greyish green and red patches; predominantly dark grayish green in lowest 0.12m	0.50
Siltstone, hard, greenish grey, blocky	0.14
Siltstone, brownish red; some mottled green patches and gypsum veinlets	0.20
Gypsum (part of nodular mass?)	0.16
Siltstone, brownish red, blocky, manly; strong green mottling in places; thin gypsum veinlets near top 1.52	
Siltstone, greyish green, with brownish red mottling; faint lamination in top 0.03m; some near-vertical fissures *Seen for*	1.38

(End of borehole)

Dating and biostratigraphy:

No ammonoids were observed in the basal Lias beds proved in R13 and these are therefore regarded as pre-Hettangian (Cope *et al.,* 1980) and are assigned a late Triassic (late Rhaetian) age.

The Mercia Mudstone and Penarth Group sequence of R8 has been examined for palynomorphs by G.W. The lowest sample, from 0.4m below the Blue Anchor Formation, proved devoid of palynomorphs. A very sparse assemblage including the miospore *Ovalipollis pseudoalatus* (Thiergart) Schuurman 1976 was recovered from 5.5m above the base of the Blue Anchor Formation.

Richer assemblages were obtained from beds higher in that formation and from the Penarth Group (Fig. 3). Those from the Blue Anchor Formation are similar in composition to assemblages documented from the unit near its type locality (Warrington and Whittaker 1984). They comprise only miospores, of land plant origin, and are dominated by *O. pseudoalatus, Rhaetipollis germanicus* Schulz 1967 and circumpolles *(Granuloperculatipollis rudis* (Klaus) Venkatachala 1966, *Corollina zwolinskai* Lund 1977, and *Classopollis spp.).* Richer assemblages characterized by associations of the miospores *O. pseudoalatus, R. germanicus, Classopollis spp.* and *Ricciisporites tuberculatus* Lundblad 1954 with organic-walled microplankton occur in the Penarth Group. Their diversity and the importance of the organic-walled microplankton component increases upwards through the sequence examined; miospores, principally *O. pseudoalatus* and *Classopollis spp.,* are dominant in the lower part of the Westbury Formation but organic-walled microplankton, principally the dinoflagellate cyst *Rhaetogonyaulax rhaetica* (Sarjeant) Loeblich and Loeblich *emend.* Harland *et al.* 1975, become abundant and are ultimately dominant in the higher beds (Fig. 3).

The palynostratigraphy of the Penarth Group succession is comparable with that documented from Westbury Formation and Cotham Member sequences elsewhere in southern Britain (Orbell 1973; Warrington 1977, 1982, 1984) though the latter unit is not identified in R8 (Fig. 2). However, assemblages comparable with those from the member elsewhere occur in the lowest 0.8m of the Lilstock Formation (Fig. 3) and indicate that it is represented within that sequence which is regarded as disturbed by cambering.

The palynomorph assemblages are indicative of a late Triassic, Rhaetian, age for the Penarth Group and the upper beds of the Blue Anchor Formation. The incoming of *Tsugaepollenites? pseudomassulae* (Mddler) Morbey 1975 at 4.45m below the Penarth Group may indicate approximately the position of the base of the Rhaetian, and beds below that level may be of late Triassic, Norian, age.

Bruton

Bruton No. 1 Borehole was drilled in 1982 for the Institute of Geological Sciences at a site (ST. 6896 3284) some 25km east-north-east of Langport in an area where the Triassic is concealed beneath younger Mesozoic rocks.

The succession proved at Bruton was briefly described by Holloway and Chadwick (1984). The Mesozoic rocks were not cored and no biostratigraphical information is available from that sequence. The succession (Fig. 4) was interpreted from a study of cuttings samples and analysis of geophysical wire-line log profiles which were related to those from the cored Burton Row Borehole, sunk on Brent Knoll some 38km to the north-west (Whittaker and Green 1983; Whittaker *et al.* 1985), and the Winterborne Kingston Borehole, drilled a similar distance to the south-south-east in Dorset (Rhys *et al.* 1982). The Langport Member of the Penarth Group generates a distinctive response on gamma ray and sonic borehole logs, and the sequence between 268.5 and 275.2m is assigned to that member. The upper and lower

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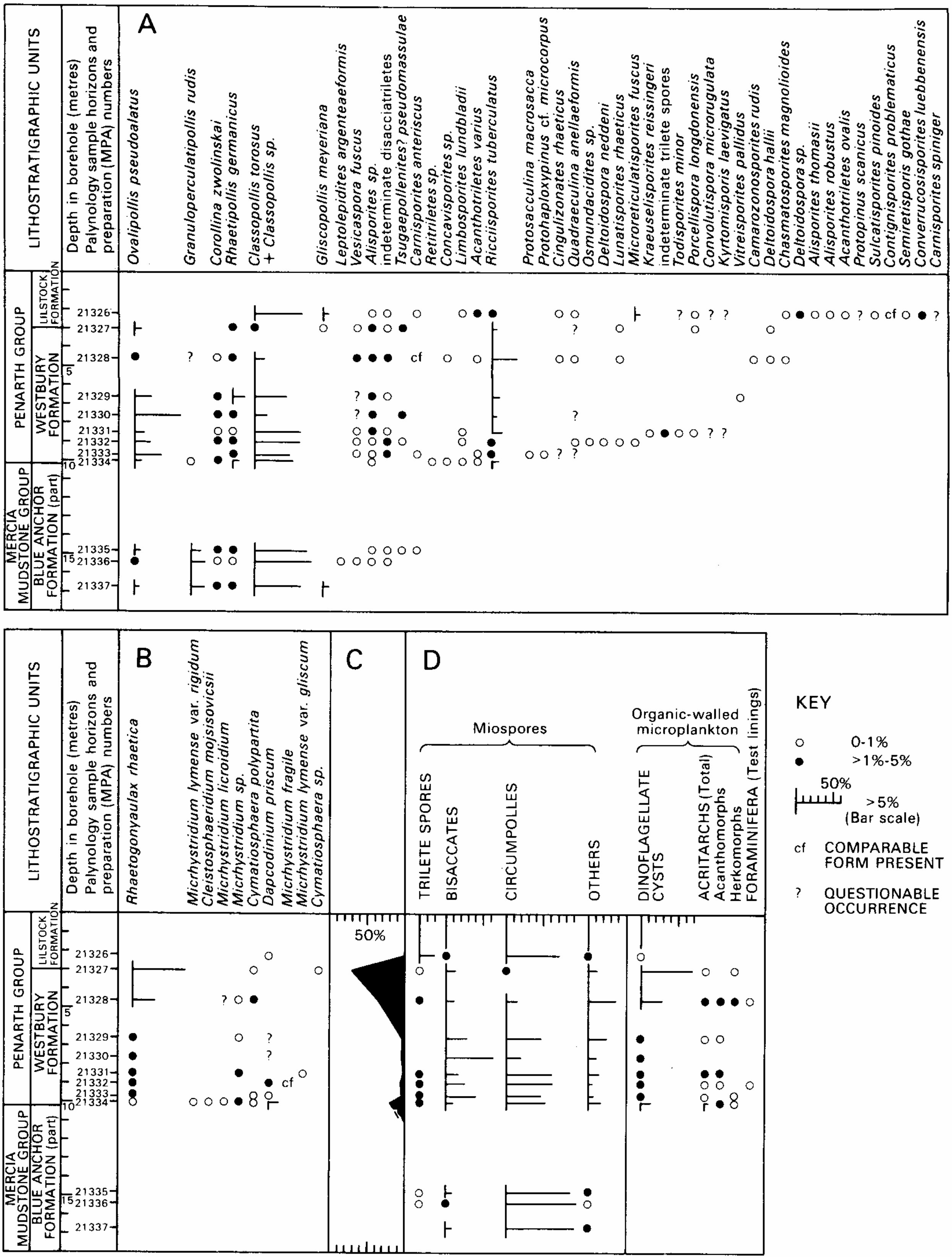

Figure 3. Distribution and relative abundances of palynomorphs from borehole R8, High Ham.
A - miospores; B - organic-walled microplankton; c - relative abundance of miospores (white) to other palynomorphs (black); D -major groups of palynomorphs.
Preparations are held in the palynological collection at the British Geological Survey, Keyworth, and are registered in the MPA series; relative abundances are based upon counts of 200 specimens.

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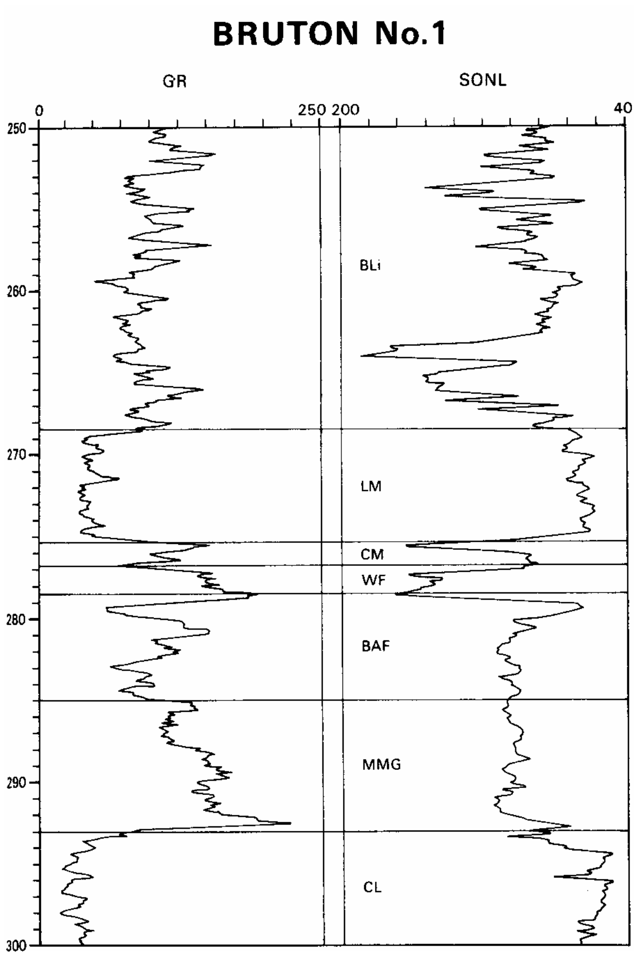

Figure 4. Gamma ray (GR) and sonic log (SONL) profiles for the 250-300m interval, Bruton No. 1 Borehole, east Somerset, from digitised records, Deep Geology Research Group, British Geological Survey, Keyworth.
BLi - Blue Lias; LM - Langport Member; CM - Cotham Member; WF - Westbury Formation; BAF - Blue Anchor Formation; MMG - red beds of Mercia Mudstone Group; CL - Carboniferous Limestone.

boundaries of the Westbury Formation are placed, respectively, at distinctive changes on the log profiles at 276.7 and 278.5m. The base of the Blue Anchor Formation is located at 285.1m, at the level of a marked change in the gamma response, and the base of the Mercia Mudstone Group, resting unconformably upon Carboniferous Limestone, is marked by a change in the gamma log profile at 293.0m.

The lithological characters of these units (Holloway and Chadwick 1984) are comparable with those described above from the sequence in the High Ham boreholes. However, marked changes in thickness (Table 1) occur between Bruton and Langport, a distance of some 25km, and between Langport and Blue Anchor on the west Somerset coast, a distance of some 35km. The thickness of the Mercia Mudstone Group present beneath the Blue Anchor Formation at Langport is unknown but is likely to be considerably greater than the 7.9m present at Bruton. At Compton Dundon (c. ST. 4832), 7.5km north-east of Langport, these beds are more than 185m thick (Moore 1867). The Blue Anchor Formation shows a marked thinning eastwards from its type locality and the Westbury Formation shows a parallel trend. The Cotham Member is remarkably uniform in thickness at the localities considered but the succeeding Langport Member displays a strong eastward thickening.

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Read at the Annual Conference of the Ussher Society, January 1986

The Jersey Main Dyke Swarm, Jersey, Channel Islands

G.J. LEES

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The Jersey Main Dyke Swarm occupies a zone some 3 km wide, whose outcrop is restricted to the extensive reef platform skirting the southern coast of the island. It cuts the SE and SW plutonic complexes. The major dyke orientation lies between NE-SW and ENE-WSW, with crustal extension values reaching 10%. The time interval between emplacement of the plutons and dyke injection in the SE was very short.

The rock types involved in the swarm are varied, comprising dolerites, microbojites, microdiorites, appinitic hornblende-lamprophyres, and feldspar-quartz porphyries.

Chemically the swarm shows a clear acid-basic bimodality, but other parameters indicate that it has a mature calc-alkaline character, characteristic of basalts generated at an active continental margin.

The Jersey Main Dyke Swarm represents the last igneous phase of the Andean - type Cadomian orogenic cycle in Jersey, intruded at a high level in the crust during a period of crustal extension in Lower Palaeozoic (Ordovician) time.

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Regional Setting

Jersey, together with the other Channel Islands, Cap de la Hague, and the northern Cotentin Peninsula constitute a late Proterozoic - earliest Phanerozoic basement in the northernmost part of the Armorican Massif dominated by the Cadomian orogenic belt. Jersey forms what seems to be a late and high level segment of this belt.

There are six major units making up the geological framework of Jersey. A late Proterozoic supracrustal sequence. - the Brioverian, is overlain by an intermediate to acid volcanic sequence, which is in turn overlain by a post -orogenic molasse deposit - the Rozel Conglomerate. Three post - Brioverian igneous complexes form the NW, SW, and SE corners of the island. The NW and SE complexes comprise both basic and acid plutonic components, while the SW complex is essentially an acidic one. The igneous complexes postdate the volcanic sequence, but their age relationship with the Rozel Conglomerate is uncertain. Whole rock Rb/Sr isochrons for these complexes have yielded dates of 565 ± 12 Ma (SW), 520 ± 4 MA (SE), and 490 ± 15 Ma (NW) (Adams, 1967a, 1976). However, more detailed recent work in Jersey (Bland, 1984) and in Guernsey (R.D'Lemos, pers. comm.) has suggested a significant downwards revision of some 50 Ma for the ages of intrusion of the younger components of these complexes. K/Ar mineral ages on amphiboles and biotites in the SE complex (Adams, 1967b) range down as low as c.440 my, suggesting this to be the time when the mobility threshold for the system in the cooling crust was passed.

The age distribution for the Jersey plutons shows them all to be very late Cadomian intrusions. Initial "Sr/ ^{16} Sr isotope ratios are between 0.7041 and 0.7085 (Adams, 1967a), suggesting that the complexes are more likely to be mantle - derived than remelts of much older granitoid crust.

The petrology and overall chemistry of Cadomian igneous rocks in the Channel Islands and N. Brittany accords well with an active continental margin model for the belt in the northern Armorican Massif. Calc-alkaline plutonic complexes related to the Cadomian orogeny form a elongated belt running WSW from Cap de la Hague in the northern Cotentin, through the Channel Islands (Alderney, Guernsey, Sark, Jersey, and possibly the Miniquiers reef), the North Tregor (Graviou and Auvray, 1985), the Petit Tr6gor (Griffiths, 1984), to the Gneiss de Brest complex in West Finistere (Adams (1967a). These complexes, where they have been investigated, have the characteristics, petrological and chemical, of I-type igneous complexes (Chappell and White, 1974) derived from mantle differentiation rather than from crustal melting. There is little evidence to suggest that these complexes were not emplaced into pre-existing continental crust. An active continental margin would seem to be a more reasonable model for this N Cadomian belt than an island arc. The location of the subduction zone to which such an active continental margin could be related is unknown. Attempts have been made (Lefort, 1977, Auvray and Lefort, in Roach, 1980)

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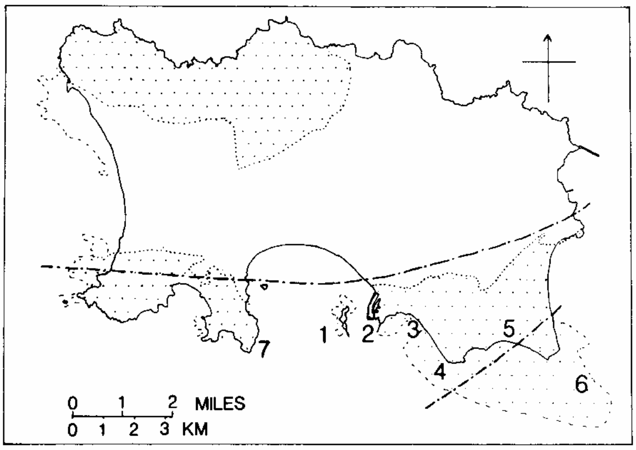

Figure 1. Simplified map of Jersey, C.I. showing the approximate boundaries of the Jersey Main Dyke Swarm between the dashed lines. The outcrops of the three plutonic complexes are marked with dotted ornamentation.

1. Elizabeth Castle; 2. La Collette; 3. Dicq Rock and Greve d'Azette; 4. Le Croc Point; 5. Pontac; 6. Seymour Tower; 7. Noirmont Point.

to identify a WSW-ENE trending magnetic anomaly, located in the English Channel some 50km to the NW, as oceanic crust related to this subduction (see Lees *et al.,* 1986, for discussion).

Minor intrusions are a prominent feature in all the Channel Islands. In Jersey they are heavily concentrated in the igneous complexes; they are sparsely distributed in the Brioverian and almost absent in the Volcanics. Using the criteria of trends and overall petrogenetic affinity, it is possible to assign nearly all the minor intrusions to one of a limited number of groups, the largest of which constitutes the Jersey Main Dyke Swarm. However, there are many dykes which cannot be directly related to these groups. Many of these have close affinities with the main dyke swarm, so may be temporally related.

Field relationships

For the detailed investigation of the dykes of SE Jersey, the large scale unpublished map prepared by Dr A.C. Bishop has been used. The larger scale features of this map have been incorporated in the 1:25000 I.G.S. Channel Islands geological sheet No. 2 (1982). Elsewhere dykes have been located using aerial photographs. Grid references given after locations in this paper refer to those used on the I.G.S. map.

The Jersey Main Dyke Swarm (hereafter JMDS) is the major expression of dyke injection to be seen in Jersey (Fig. 1). Exposure of this swarm is restricted to the coastal platform which skirts the southern coast of the island - essentially a dissected 8m raised beach. The platform is most extensively developed off the SE coast, where reefs up to 2km offshore are exposed during the mid to low range of a 11.6m spring tide.

The main dyke swarm appears to occupy a rather narrow belt some 3km wide where it is fully exposed in the SE complex between Elizabeth Castle (648476) and Pontac (694468) (Fig. 1). Here it is trending in a NE-SW to ENE-WSW direction (the modes are 045° and 060°) (Fig. 2). Local crustal extension values up to c.10% may be found. In south-west Jersey, only the northern margin of the swarm is visible because the overall trend has swung round to E-W (the mode is 085°) (Figs. 1 and 2). There is much evidence for local joint control on the direction of dyke injection on a small scale. This control accounts for a large part of the spread seen in Fig. 2.

In SW Jersey, there is much less variety in dyke rock type than is seen in the SE of the island. W of Noirmont (Fig. 1), only metadolerite dykes are seen. However, no difference can be discerned between the compositions of these dykes and those further E.

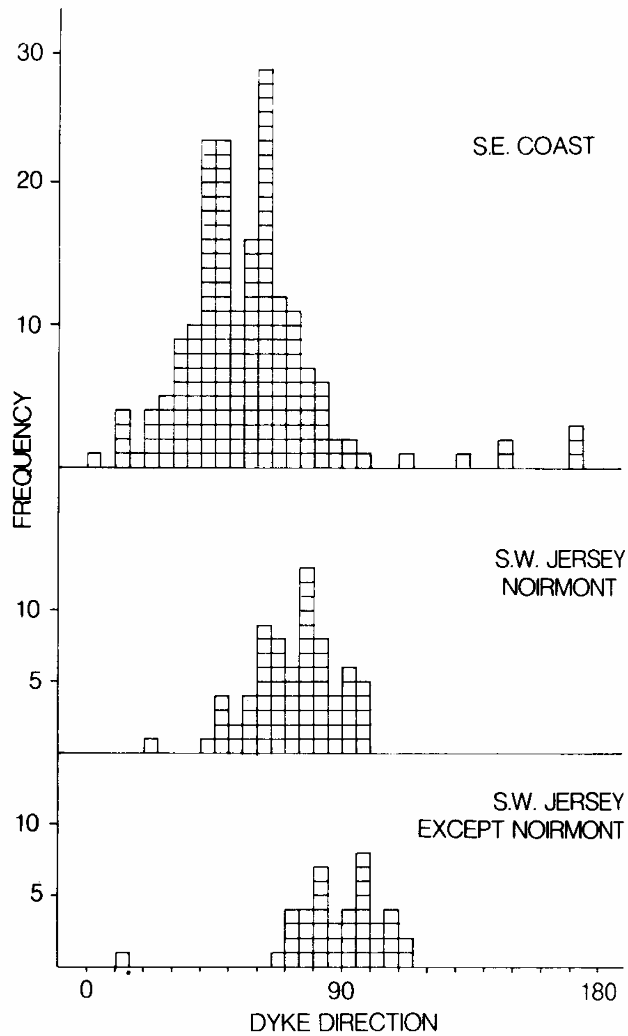

Figure 2. Histogram of dyke orientation in the Jersey Main Dyke Swarm in three zones: SE Jersey; SW Jersey in the Noirmont peninsula; SW Jersey excluding the Noirmont peninsula. The change in the modal direction of the swarm is apparent.

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The most intensive development of the JMDS occurs in SE Jersey. The SE plutonic Complex comprises a suite of older granitoids, cropping out in the W and NE of the exposure, which intrude a melanocratic diorite body to form an agmatite complex. This agmatite complex is in turn intruded by a Newer granite (the Le Hocq Granite) in the centre of the complex, to form another later agmatite zone along its contact, while extensive veining of the older agmatite also occurs. In this complex, off the Greve d'Azette (665470), early basic members of the JMDS are cut across by these Newer granite veins.

Ample evidence exists elsewhere in the SE complex to confirm that the time interval between the emplacement of the last major phase of the plutonic complex - the Le Hocq Granite - and the injection of the dyke swarm was a very short one. Basic dykes contain many armoured quartz xenocrysts and are cut in places by thin aplite sheets up to 0.1 m thick, eg. at Pontac (694468). Acid veins penetrate the basic margins of composite dykes in places and small-scale marginal mobilization of the country rock may smetimes be seen. Near Seymour Tower (726457), a basic sheet cutting the Le Hocq granite has been emplaced seemingly as a series of boat-shape-ending pods, recalling syn-plutonic dyke injection elsewhere (eg. Gobbing and Pitcher, 1972).

A fairly wide range of rock types is encountered in the intrusions of the JMDS. Numerically, basic dykes predominate, but some having more intermediate compositions also occur. A volumetrically significant component is acidic in composition, often forming the leucocratic centres of the relatively thick composite intrusions.

Local injection sequences may be erected for different parts of the main swarm. However, the absence of any ubiquitous and distinctive composition makes it impossible to construct an overall scheme covering the whole outcrop of the swarm. More detail of these sequences may be found in Bisson and Bishop (in press).

The overall character of composite minor intrusions has been reviewed by Blake et al. (1965). The presence of such composite intrusions is usually taken to indicate the contemporaneous presence of magmas of contrasting composition. They tend to be confined to the central zone of the swarm, and occur throughout the timespan of eruption, though the compositional contrast shifts from an acid to intermediate one in the earlier intrusions (eg. La Collette sill, Le Croc `appinite' dyke), to an acid to basic one in the later dykes. These composite dykes tend to be much thicker than the simple basic ones, ranging from a metre or two up to 10m. The basic margins of such dykes, however, tend to have a constant thickness (Fig. 3). Virtually all variations may be found between simple basic dykes, through dykes with thin central acid screens (eg. dyke I in Fig. 4, and dyke at Elizabeth Castle (648476)), to dykes in which the basic margins are of insignificant thickness compared with the acid centre

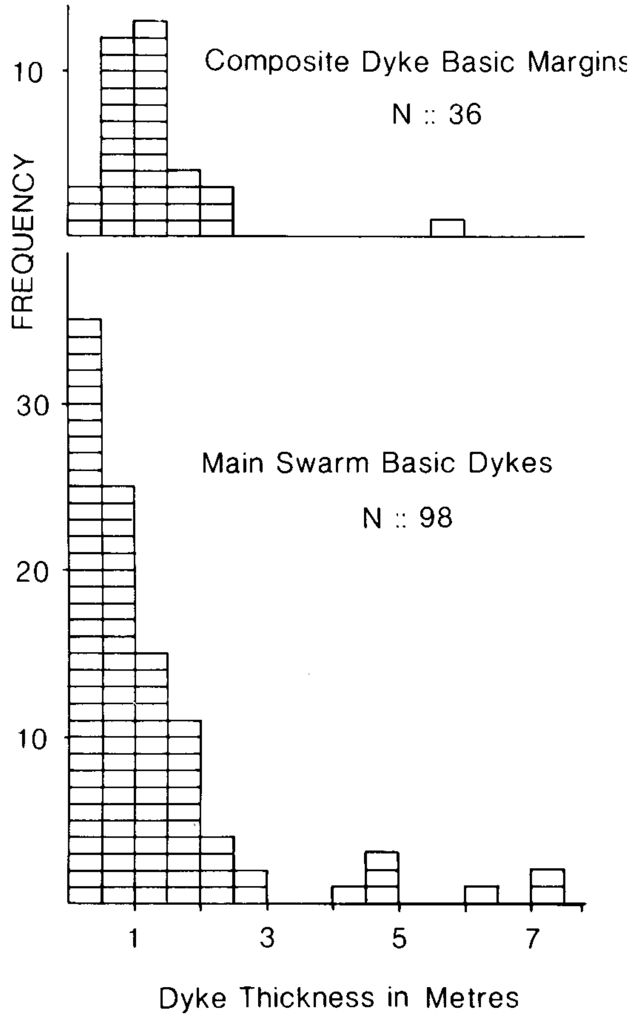

Figure 3. (a). Histogram of the thickness of basic and intermediate dykes in metres.

(b). Histogram of the combined thickness of the basic margins of composite intrusions.

(dykes 2 and 3 in Fig. 4), and then to dykes in which they become intermittent (dyke 5 in Fig. 4), and finally disappear completely (Green Island (674495)).

In one large early composite intrusion running seawards from Le Croc Point (673462) (dyke 4 in Fig. 4), the acid portion is confined to two screens just inside the fine-grained margins (as is the case with the La Collette sill). The centre of the dyke is occupied with 'appinite', whose contact with the porphyry is lobed and pillowed. In this pillowed zone, the intermediate rock forms ball-like masses, often with fine-grained and crenulate margins, in a net-vein-like complex of porphyry (cf. Blake et al., op cit.). It can extend right across the centre zone of the dyke. The contacts between the basic margins and the acid centres in the composite dykes are usually sharp and planar, sometimes showing chilling or pseudochilling of the basic material against the acid centre, but often not. In some cases gradational contacts may be seen.

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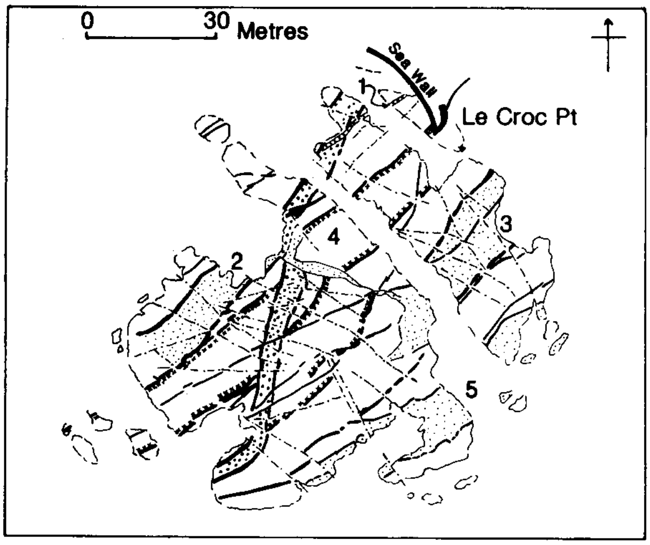

Figure 4. Simplified geological map of Le Croc Point, SE Jersey (673462), showing the cross-cutting relationships of composite dykes of the Jersey Main Dyke Swarm. Mapping is by Dr A.C. Bishop. Many minor basic dykes have been omitted for clarity.

These large composite intrusions may be seen intersecting one another; most spectacularly this occurs immediately below Le Croc Point, where three large dykes and numerous smaller ones are seen to intersect (Fig. 4).

It is tempting to use these large, easily traced bodies to erect an eruptive chronology, separating the injection sequence into a number of phases. Such a division must remain essentially artificial, however, unless evidence is available to suggest an intervening change in conditions, such as a change in dominant stress direction, or injection of plutonic material. It is probably much more realistic to consider the process as one of continuous but sporadic intrusion into the same stress regime over a relatively short time interval (c.2-3 my. is the time span estimated for Mull (Mussett et al., 1980)).

There is evidence that, during the existence of the fracture zone into which the dyke swarm was injected, the stress field changed sufficiently to enable emplacement of dykes along N-S trends more than once, though these are not numerous enough to be useful chronological discriminators. The N-S dykes cannot be distinguished by any other character from the JMDS. The main dyke swarm injection episode thus shows a multistage history with alternating main trend phases and minor N-S trend phases, which latter include the last dykes injected.

Most of the dykes making up the swarm are fairly narrow, mostly less than lm thick, as can be seen in Figure 3. The thickness of the basic margins of composite dykes shows a very similar distribution, reinforcing the view that they do not differ from normal basic dykes in the swarm.

Petrology

The basic members of the dyke swarm are mostly doleritic or metadoleritic in composition. Both aphyric and porphyritic dykes are common, the latter being essentially plagiophyric, though pyroxene- and hornblende-phyric types may also be found. A significant number of dykes of basic composition occur in which the primary mafic phase appears to have been brown hornblende. These primary hornblende + plagioclase dyke rocks are often referred to as diorites (Bishop and Bisson, in press). All plagioclases have been modified by metamorphism so it is almost impossible to find any remaining with compositions more Ca-rich than andesine. However, these dykes have a chemical composition which is thoroughly basic (usually olivine-normative); primary modal quartz is never seen; abundant epidote is intimately associated with altered plagioclase; the normative plagioclase composition lies between An_{50} and An_{70} for nearly all the basic rocks, whether clinopyroxene- or amphibole-bearing. It is thus questionable whether most of these primary hornblendic rocks are really diorites or are strictly micro-bojites. One such spectacular example, some 10m thick, occurs between Le Croc (673462), Le Nez (678462), and Green Island (674459), but most of the basic dykes found to the east of Dicq Rock (659477) appear to have had brown hornblende as the sole or dominant mafic phase. In general it may be said that the true dolerites of the main swarm are restricted to the zone west of Dicq Rock (659477). They, together with the late N-S dykes, appear to have been emplaced late in the sequence, perhaps when the geothermal gradient had become less steep.

The intermediate members of the swarm comprise appinitic hornblende lamprophyres (?camptonites), characterised by long bladed brown hornblendes, often cored, and acicular apatites (cf. Wells and Bishop, 1955); and 'andesites' - plagioclase rich rocks with a trachytic texture. These intermediate types, together with some of the more evolved basic compositions, eg. the La Collette Sill (652476) (Bishop, 1964b), tend to be emplaced early in the injection sequence.

The acidic members of the swarm may best be described as feldspar-quartz porphyries. They comprise K-feldspar which is usually perthitic; plagioclase, almost invariably in glomeritic groups; and quartz, often embayed and corroded; as phenocrystic components set in a groundmass of devitrified rhyolitic glass, which may show a felsitic, micrographic, spherulitic, or micrographic texture, and often all four in patches. The spherulitic intergrowths usually occur as haloes around phenocrysts. These porphyries should strictly be termed porphyritic rhyolites. The markedly porphyritic nature of the porphyries affords a useful criterion for distinguishing them from veins of Newer Granite.

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Metamorphism

All dykes of the JMDS have been metamorphosed to some degree. Prehnite and pumpellyite occur in nearly all basic rocks, but is often difficult to find. Many of the dykes have been metamorphosed in the lower greenschist facies. The metamorphism appears to have been essentially static in nature, as relict igneous textures and relict primary minerals are nearly always preserved. Plagioclases retain good tabular outlines although usually totally replaced by albite, with patches of epidote, chlorite, and prehnite. Mafic minerals, especially pyroxenes, are replaced by green amphibole, while primary brown amphiboles show partial replacement and overgrowth by actinolite. Ore grains often show partial replacement by leucoxene and sphene. Vesicles are usually infilled with aggregates of calcite, chlorite, epidote, pyrite, and occasionally actinolite. The assemblage for most basic rocks can thus be said to be actinolite + chlorite + epidote + albite + prehnite, indicative of the lower greenschist facies. Dykes showing only the effects of prehnite-pumpellyite facies metamorphism can often be shown to have been emplaced late in the history of the swarm. They are those found mostly west of Dicq Rock (659477).

Geochemistry

A detailed discussion of the geochemical character and relationships of the JMDS has been held over for publication elsewhere. All that will be attempted here will be to show the overall chemical affinities of the swarm. All the rocks of the JMDS have been metamorphosed. It is not considered that metamorphism has caused changes sufficient to affect any conclusions reached from a study of the chemical character of the dyke swarm.

A histogram of SiO_2 concentrations (Fig. 5) shows a distinctly bimodal distribution. Most significant is the almost total absence of rocks with SiO_2 contents between 58% and 68%, separating the basic and intermediate members of the swarm from the acid porphyries. The distribution of the basic to intermediate components shows a mode at c.46%, reflecting the basaltic character of most of the dykes. The bimodality suggests that care must be exercised in interpreting other variation diagrams. The A-F-M diagram shows an apparently classic calc-alkaline trend, but once again a prominent compositional hiatus occurs between the intermediate and acid components of the swarm. If the basic and intermediate members alone are considered, the trend is much less well defined. However, though a slight iron enrichment trend may be detected, nearly all points plot in the CA field as delineated by Kuno (1968). The Total Alkalis vs SiO_2 plot (Kuno, 1968) shows the suite to be transitional between high-alumina basalts (calc-alkaline) and alkaline ones. This probably reflects the enhanced K_2O content. The Alkali-Lime index of Peacock (1931) for the main dyke swarm is 59.8, thus falling in the calc-alkaline range. Using the ratio FeOtot/MgO as a differentiation index (Miyashiro, 1974), the basic and intermediate members of the swarm show a clear SiO2 enrichment trend. The relatively low TiO_2 contents are characteristic of subduction related basalts.

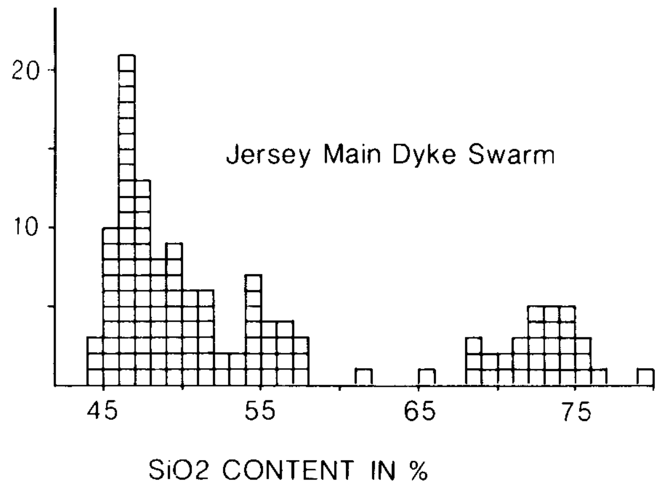

Figure 5. Histogram of the SiO_2 content of 121 samples of the Jersey main dyke swarm, showing its bimodal nature.

Whitford *et aL* (1979) have used the K_2O - SiO_2 relationship to subdivide the CA series in the Sunda Arc, Indonesia. This classification places most of the basic rocks of the Jersey swarm as High-K calc-alkaline, though transitional into normal calc-alkaline.

The C.I.P.W. norms for the basic rocks of the JMDS, after normalizing the Fe_2O_3/FeO ratios to 0.3 and ruling out samples with excess LOI values, show nearly all to be olivine-normative. A certain number are quartz-normative; these are usually the more evolved rock types and include the 'appinites'.

The binary plot of Y vs. Zr shows the classic calc-alkaline distribution, ie. a positive correlation but very limited range of Y for a rather wider range of Zr. Most basic and intermediate rocks lie between 17 and 35 ppm Y and 60 and 190 ppm Zr, though a few reach c. 45 ppm Y and c.300 ppm Zr. The acid rocks show a different distribution, being much lower in Y for equivalent Zr contents.

The chondrite-normalised REE profiles of a selected group of samples from the main swarm may be seen in Figure 6. The basic and intermediate components (Fig. 6a) display a remarkably parallel set of patterns. They show fairly high LREE enrichment (up to 150 times chondritic); the slightly concave upward patterns show a point of inflection about Dy. There is some HREE depletion relative to MORB. No significant Eu anomalies are visible. Such patterns are characteristic of calcalkaline basalts (Hanson, 1980).

The REE profiles of the acid components of the swarm (Fig. 6b) show significantly different patterns to those of

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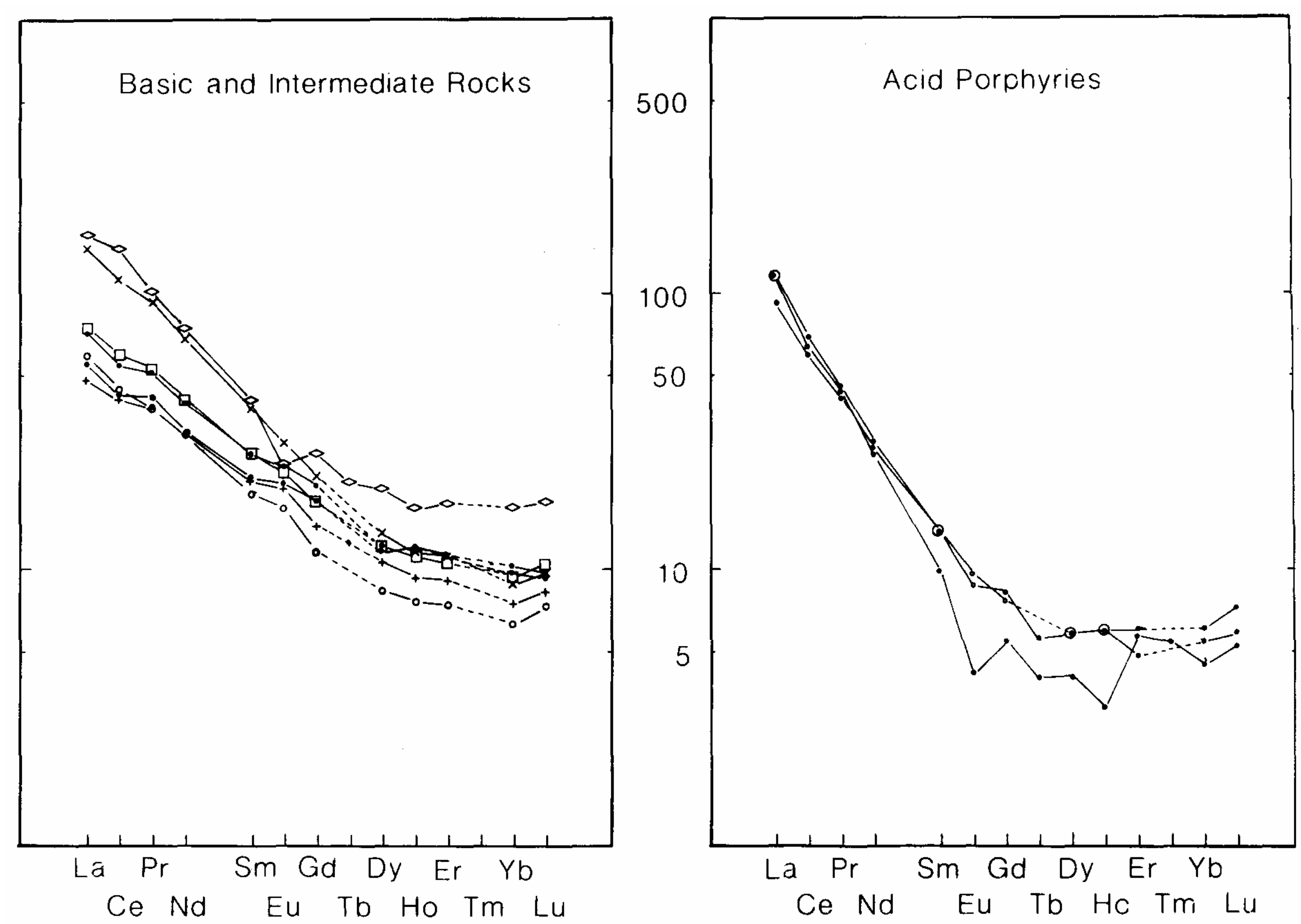

Figure 6. Chondrite normalized REE profiles (normalization values of Nakamura, 1974). (a) Profiles of basic and intermediate rocks. (b) Profiles of acid porphyries.

the basic and intermediate rocks. The acid porphyries show similar levels of LREE enrichment, but much lower HREE values. Thus, their patterns, while still concave upward, show a much more marked point of inflection about Eu and Gd. Again, no significant Eu anomaly is seen.

These patterns would confirm that the basic and acid members of the swarm cannot be related by simple high level (ie. low pressure) fractionation of basaltic magma *(sensu. Cox,* 1980).

The overall consensus of the geochemical characteristics of the basic and intermediate rocks of the JMDS is that of a suite of calc-alkaline basalts of high-K type. The acidic members of the swarm forma distinct group. It may be possible to produce such a discontiguous group by fractionation of some mineral species like amphibole, but the possible alternative of producing it by partial melting of granitic crust cannot be ruled out (cf. Cobbing and Pitcher, 1972). Figure 7 shows the envelope of profiles formed by 5 typical main swarm dolerites, when plotted on a MORE-normalized incompatible element diagram (spidergram) using the normalisation scheme of Pearce (1983). Spidergrams of this sort can be very useful in suggesting possible geotectonic environments for eruption. The pronounced "hump" caused by enrichment of the LIL elements Sr to Th clearly indicates that these are subduction related basic rocks. The significant enrichment of the coherent element pair Nb-Ta relative to that of the pair of Zr-Hf, and of both relative to the Pair Y-Yb is readily seen. This would suggest that the magma was being derived from a source enriched in HFS elements. There is no evidence to suggest that pre-existing continental crust did not underlie Jersey during the Cadomian orogenic episode, so these features would suggest quite strongly that the JMDS was emplaced in an active continental environment underlain by enriched sub-continental lithospheric mantle in the supra-subduction zone slab (see Pearce, 1983, for discussion).

The main swarm dyke basalts show patterns very similar to those obtained from Central Andean basalts.

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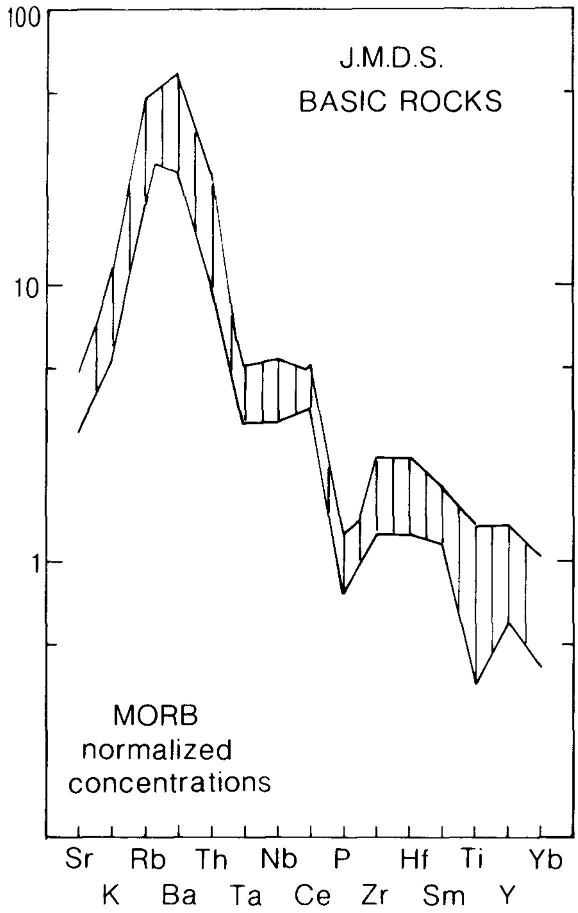

Figure 7. MORE normalized `spidergram' showing the envelope of 5 basic rocks of the Jersey Main Dyke Swarm.

Conclusions

The plutonic complexes of Jersey were among the last to be emplaced in the Cadomian cycle. No evidence now remains of their volcanic edifice, since the volcanic sequence found on Jersey predates the complexes.

The Jersey main dyke swarm in turn postdates the emplacement of these complexes, albeit by a very small time interval and even with some overlap in SE Jersey. It occupies a relatively narrow zone some 3km wide, trending between NE-SW and E-W. Crustal extension within the zone reaches c. 10%.

Chemically the dyke swarm has a calc-alkaline affinity. Detailed comparison reveals similarities more with active continental margin basalts emplaced through continental crust than with those of island arcs. The swarm is thus related to the preceding Cadomian orogenic cycle rather than to any new geotectonic regime.

The Jersey Main Dyke Swarm may be said to represent the last major Cadomian igneous cycle to be found on

Jersey, and probably in the Channel Islands as a whole. The orientation of the swarm, parallel to the length of the Cadomian orogenic belt in the Armorican Massif represents extension, possibly flexural in nature. This may be related to changes in the stress regime of the presumed adjacent subduction zone.

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Read at the Annual Conference of the Ussher Society, January 1986

The geological significance of some geophysical anomalies in western Somerset

J.D. CORNWELL

Cornwell, *J.D. 1986.* The geological significance of some geophysical anomalies in western Somerset. *Proceedings of the Ussher Society, 6, 383-388.*

The regional aeromagnetic and Bouguer anomaly maps of western Somerset reveal several significant anomalies, some of which are obviously related to known geological structures, or their extensions, while others appear to be caused by structures for which there *is* little evidence at the surface. The interpretation of the Bouguer anomaly data is complicated by the variability and range of the known rock densities - while there are comparatively few magnetic rocks known in the area which are likely to give rise to extensive aeromagnetic anomalies. Several prominent ESE-trending lineaments are recognisable from the geophysical data for the areas of Palaeozoic rocks, including the Exmoor Bouguer anomaly gradient zone; and it is suggested that these are related to faults affecting the Mesozoic rocks to the east.

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Introduction

The west Somerset area has been primarily of interest to geophysicists as a result of Falcon's observation (in discussion to Cook and Thirlaway 1952) that a Bouguer gravity anomaly gradient across the Quantock Hills might provide evidence for a large concealed thrust fault first proposed by Ussher (1908) in the Cannington Park area. Subsequent studies led to much speculation on the nature and significance, in the context of the 'Variscan Front' in SW England, of the structure indicated by the gravity evidence (Bott and others, 1958, Bott and Scott, 1964, Brooks and Thompson, 1973, Brooks and others, 1977).

An examination of the regional geophysical data (mainly Institute of Geological Sciences, 1981 and British Geological Survey, 1986) and anomalies occurring in two areas of west Somerset (Cornwell in Whittaker and Green, 1983 and in Edmonds and Williams, 1985) have provided more information on the Exmoor anomaly and also indicate the existence of several other large structures.

The possible significance of these to the geology of an area including the Quantock Hills and adjoining parts of Exmoor and the Central Somerset Basin is discussed.

Physical properties of rocks

The interpretation of geophysical anomalies requires knowledge of the physical properties of the main rock types and this is particularly important in the case of gravity surveys. For the Bristol Channel area, the densities of the main rock units (ranging in age from Silurian to Jurassic) have been measured for several studies and are summarised in the two memoirs.

Amongst the more interesting and unexpected results are the unusually high densities for the Permian to Jurassic sedimentary rocks encountered in the Burton Row Borehole sited in the Central Somerset Basin (Fig 1). The densities reflect the low porosities of the rocks and are a result of greater than expected depths of burial at some stage in the evolution of the basin and/or high geothermal gradients. Cornford (this volume) has discussed the geochemical evidence for these possibilities as they affect Liassic rocks. There is some evidence from the Taunton area (Cornwell in Edmonds and Williams, 1985) and the Bruton Borehole (Holloway and Chadwick, 1984) that the low porosity rocks are largely confined to the western part of the Central Somerset Basin and possibly beneath the Bristol Channel and into South Wales. The existence of these rocks adds to the uncertainties in the interpretation of gravity anomalies, particularly as the densities are similar to those recorded for Devonian rocks which probably form a major part of the pre-New Red Sandstone basement in the area.

Igneous rocks are the most likely sources of aeromagnetic anomalies and lavas of both Silurian and Carboniferous age in the area are known to contain magnetite. The former give rise to significant anomalies on the aeromagnetic map (Brooks in Green and Welch 1965) while the more restricted Carboniferous volcanism is only apparent geophysically from the results of ground magnetic surveys (Kearey and Rainsford, 1979).

Interpretation of geophysical anomalies

1. Central Somerset Basin

Containing a thickness of Mesozoic sediments in excess of the 1.1km proved in the Burton Row Borehole and

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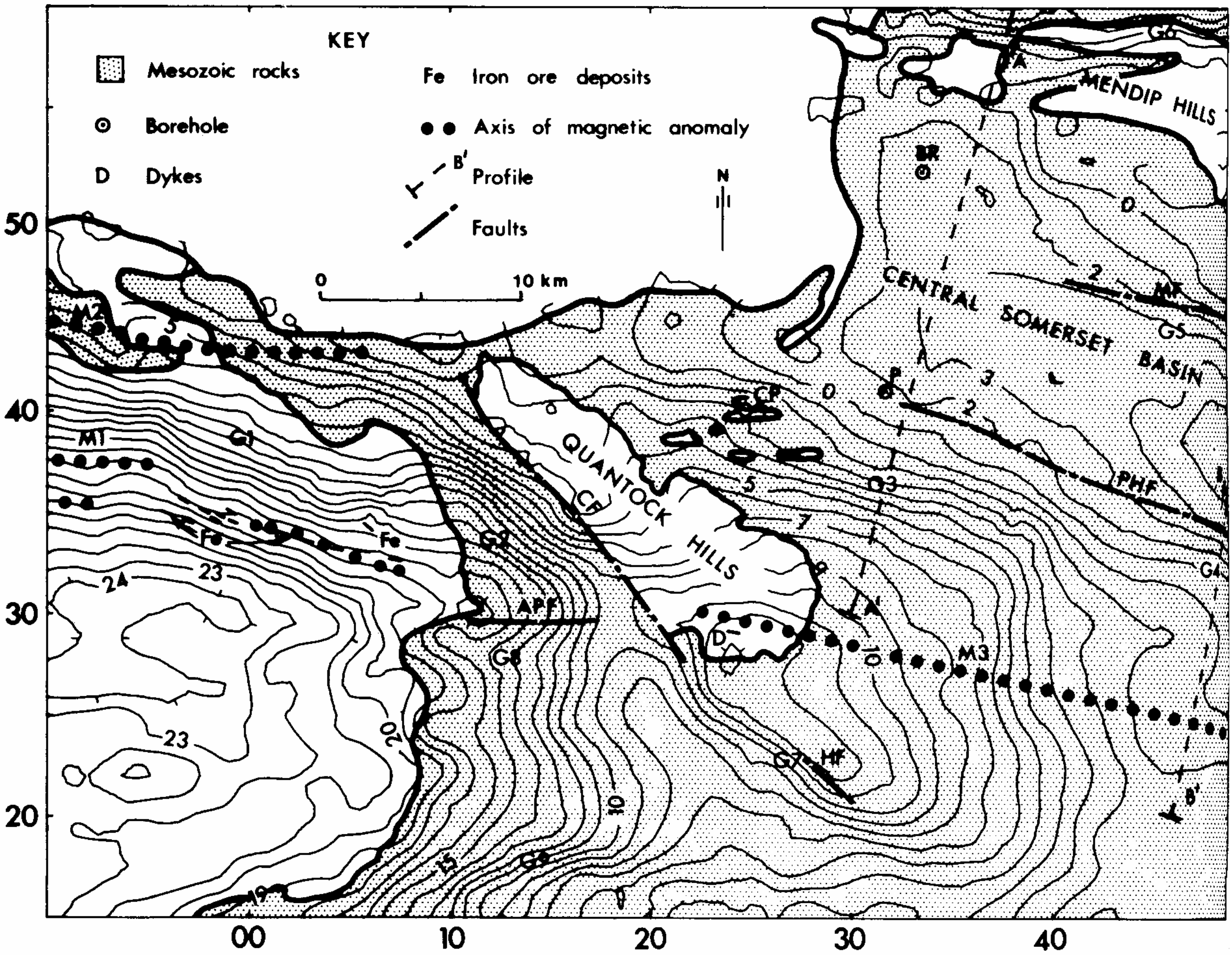

Figure 1. Bouguer anomaly map (with contours as 1 mGal intervals) of the west Somerset area with locations of the geophysical anomalies and selected geological features. The numbered geophysical anomalies are referred to in the text. Key to abbreviations: **faults:** APF - Ash Priors, HF - Hatch, MF - Midgley, PHF - Polden Hills, CF - Cothelstone; **boreholes:** BR - Burton Row, CP - Cannington Park (Knap Farm), P - Puriton.

probably resting on a basement of Carboniferous and Devonian rocks, the Central Somerset Basin would be expected to give rise to a large Bouguer anomaly low. The presence near the coast of a poorly defined low can be explained by the high densities of the Mesozoic rocks in the Burton Row Borehole and probably also by a regional northward decrease of Bouguer anomaly values due to density changes at a deep level in the crust. The theoretical gravity response of the proven sequence is relatively small (Fig. 2a) and in particular fails to explain the residual anomaly at Puriton where another low density body of unknown origin has to be postulated.

Further to the east there is a more pronounced Bouguer anomaly low associated with the Central Somerset Basin and the observed profile shown in Figure 2b can conveniently be explained by a graben filled with low density Mesozoic rocks and confined by the Polden Hills and Mudgley faults. The lower densities used in modelling the gravity data in this area not only produce an acceptable model but their existence is to some extent indicated by the results from the Bruton Borehole *where* the geophysical log (Holloway and Chadwick, 1984) shows an average density for the Mesozoic rocks of 2.41 Mg/m^3 compared with the average of 2.61 Mg/m^3 at Burton Row (although the sequences differ in the two boreholes). It could also be argued that because the basement high proven at Bruton is associated with a distinct Bouguer anomaly then significant density contrasts must exist. An alternative interpretation of the profile in Figure 2b would require an improbably great thickness of low density rocks of unknown nature at depth between the two faults shown. The geophysical evidence therefore suggests that the evolutionary history of the Central Somerset Basin may have differed significantly in the two areas.

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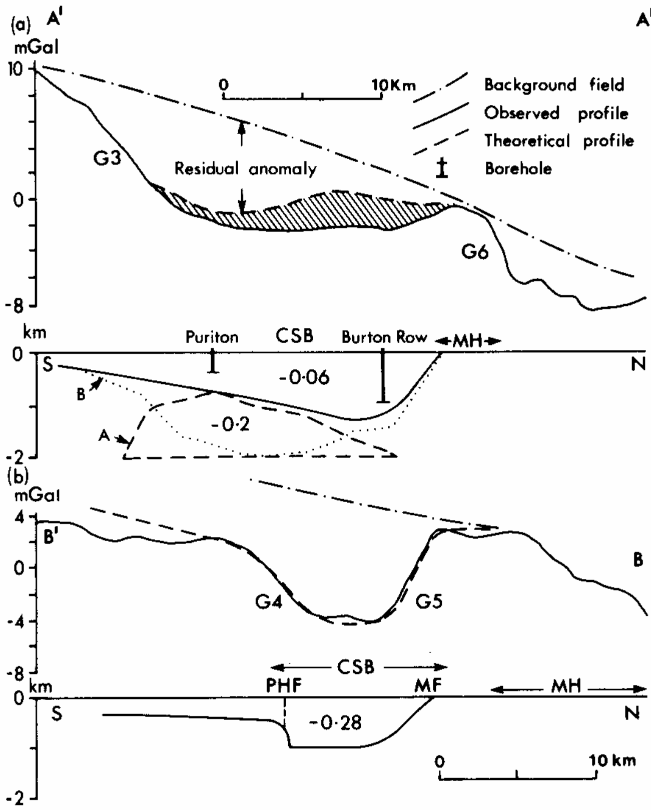

Figure 2. Bouguer anomaly profiles AA' and BB' (see Fig. 1 for locations) across the Central Somerset Basin. In Fig. 2a the shaded area represents the computed gravity effect of the Mesozoic rocks with an average density (2.63 Mg/m^3) similar to that in the Burton Row Borehole. The residual anomaly can be explained by either model A or B. In Fig. 2b the model shown has an average density similar to that in the Bruton Borehole. Key to abbreviations MH - Mendip Hills, CSB - Central Somerset Basin.

The Bouguer anomaly values on the north side of the profile in Figure 2b level off over a concealed shelf of shallow Palaeozoic rocks south of the Mendip Hills (Brooks in Green and Welch, 1965).

II. Exmoor

The Bouguer anomaly gradient mentioned earlier is expressed on the gravity map as a group of regularly ESE-trending contours with decreasing values towards the Bristol Channel (G I Fig. 1). The low density body which must exist to explain the anomaly has been interpreted as a wedge of low density Devonian or Carboniferous sediments beneath a thrust block formed by the Devonian rocks of Exmoor (Bost and others, 1958) and as the Lower Palaeozoic or Precambrian core of the north Devon anticline (Brooks and others, 1977). Although the explanation of the anomaly remains uncertain more detailed regional gravity survey data suggest that part, at least, is due to the southward thickening of the higher density members of the Devonian sequence, notably the Ilfracombe and Morte Slate units. This conclusion is based partly on the observation that the Bouguer anomaly values increase abruptly at the base of the Ilfracombe Slates and decrease at the top of the Morte Slates. The succeeding horizon, the Pickwell Down Sandstones, coincides with almost level Bouguer anomaly values or even in places with distinct lows, a feature ascribed by Al-Sadi (1967) to the fact that these arenaceous rocks are about 0.12Mg/m3' less dense than the slates.

The contours indicating the Exmoor gradient zone, when traced eastwards, change from having an ESE-trend to a more south-easterly trend just to the west of the Cothelstone Fault (G2 Fig. 1) and are then apparently displaced southwards over the Quantock Hills. This would be consistent with an abrupt change in the strike of the Morte Slates in the area concealed by Permo-Triassic rocks just to the east of the Brendon Hills and, in the Quantock Hills, with the outcrop pattern in the Quantock Anticline. The observed arrangement of anomalies therefore seems to favour an interpretation in this area which is closely related to the surface geology as it is more difficult to explain sudden changes in the direction of a deep-seated thrust plane. The possibility that the gradient zone extends east of the Quantock Hills is discussed later.

The generally smooth aeromagnetic map of west Somerset and north Devon is interrupted by two linear ESE-trending anomalies. One of these (M1 in Fig. 1) is very pronounced and indicates a near-surface magnetic horizon, 60km long, near the top of the Morte Slates; recent drilling evidence led Tombs (in Edmonds and others, 1985) to interpret this as a pyrrhotite-bearing mineralised horizon lying parallel with the southward dipping bedding planes of the slates. The second, more northerly anomaly (M2 Fig. 1) is less well-defined and suggestive of a more deep seated body - again striking to the ESE. The approximate coincidence of this anomaly with the outcrops of the Hangman Grits and Lynton Beds suggests that it might be due to concealed magnetic rocks in the core of the north Devon anticline at estimated depths of about 2-3km.

III. The Quantock Hills and their extensions

The Quantock Hills coincide with a gradual southward increase in anomaly values which culminates just south of the exposed Devonian rocks to form a well defined high. This has the appearance of being due to a horst structure probably composed of Pickwell Down Sandstones and younger sediments, beneath the Permo-Triassic rocks. The south-western margin of this high is particularly well-defined by a zone of steep Bouguer anomaly gradients (G7 Fig. 1) due to a density boundary which cannot be deeper than about lkm and probably represents the thickening, by at least 0.6km, of Permo-Triassic rocks to the SW. The zone continues the line of the Cothelstone Fault and provides convincing evidence that this structure links up with the Hatch Fault (Fig. 1),

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as suggested by Whittaker (1972) on the basis of surface mapping.

The main magnetic feature in the Quantock Hills area is a low amplitude anomaly elongated towards the ESE (M3 Fig. 1) which coincides with an elongated ridge in the pre-Permian basement surface.

Discussion: the regional significance of the geophysical anomalies

Several of the most significant anomalies described above trend to the ESE, parallel with the structural grain of this and many other areas in this part of SW England. It is proposed that the geophysical evidence suggests the existence of at least three major lineaments which could continue eastwards into the Wessex Basin.

The Exmoor gravity gradient zone extends for a distance of at least 70km before being deflected near the Cothelstone Fault. East of the Quantock Hills another gradient zone continues the line of the Exmoor anomaly through Cannington Park and merges with a step-like anomaly over the Polden Hills Fault (G4 Fig. 1); it seems likely that these features have a common cause. One implication of this however is that as the gradient has been tentatively assigned to the southward thickening of Devonian rocks, particularly the slate members of the sequence, then the same geological structure that exists on Exmoor must reappear at depth east of the Quantock Hills. This can only be achieved if the Quantock Anticline is followed immediately to the east by a tight syncline or extensive faulting so that a general ESE strike is restored and continues into the Central Somerset Basin.

The Exmoor anomaly caused by the pyrrhotite Mineral-isation is a remarkable linear feature running parallel with the Bouguer anomaly gradient zone with a direction of 110° E of N. Extrapolating the line of the magnetic anomalies eastwards the trend and position is repeated by the magnetic anomaly on the SE side of the Quantock Hills (M3 Fig. 1). This overall alignment could be dismissed as fortuitous were it not for the occurrence in the gap between the two magnetic anomalies of two other apparently unrelated geological features - the isolated group of dykes at Hestercombe in the south part of the Quantock Hills (D Fig. 1) and, in the Brendon Hills, the group of iron-ore bearing veins in the Morte Slates (Fe in Fig. 1). The igneous rocks are clearly intrusive (Evens and Wallis, 1930, Edmonds and Williams, 1985), with an age of 264±36 Ma, while the veins probably have a hydrothermal origin (Scrivener and Bennett, 1980). Both the veins and the intrusions are aligned parallel with the bedding of the adjacent sediments; in the former case the alignment forms part of a linear regional pattern but the dykes occur where the strike of the sediments is only locally to the ESE near the axis of the Quantock Anticline. The alignment of all these features tends to suggest a common structural

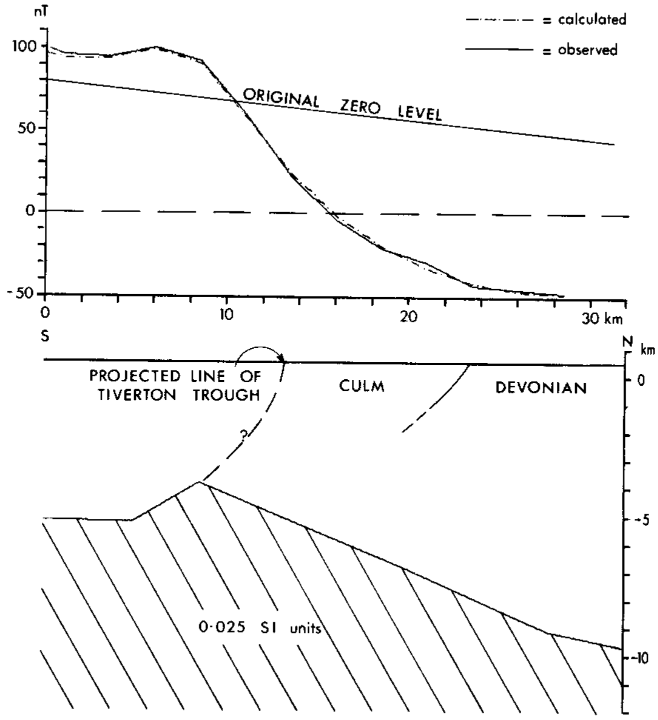

Figure 3. Magnetic profile CC' (see Fig. 4 for location) and a model for the magnetic basement.

control and it is suggested they could have been formed in a tensional fracture system, which gave access at various localities to mineralising fluids and to a basic magma.

The pronounced ESE trend, which is such a characteristic feature of these two geophysical lineaments, is repeated to the south by a step-like change in total magnetic field values (Fig. 3 and M4 Fig. 4), with the smoothness of the contours indicating a deep-seated origin of 4 to 5km. The form of the anomaly can be modelled as a magnetic body, rising southwards to a peak and then descending again towards the Dartmoor Granite; further south the anomaly is affected by the granite bodies. The anomaly terminates to the east near the margin of the New Red Sandstone outcrop in the vicinity of the Tiverton Trough but westwards it seems to reappear the other side of the magnetic field anomalies associated with the Lundy intrusions.

The geophysically defined lineaments described above all have considerable extent horizontally and, in most cases, vertically. Although their trend coincides with that of the regional strike of the sediments it is easier to accept the lineaments as representing fault planes because of their extent and linear character. In the case of the Exmoor Bouguer anomaly gradient, possible contributions due to lithological variations within the Devonian sequence and between this and the basement have been discussed but it is likely that this has been accentuated by strike faulting, which was reactivated to give rise to the Polden Hills Fault in the Mesozoic sediments and probably the faulting separating the Carboniferous inliers around Cannington from the adjacent Devonian inliers. A fault

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coincident with the Exmoor magnetic lineament and formed during a tensional phase seems necessary to explain the linear nature of disparate geological events. The comparatively small angles (about 10° to 30°) of the deep-seated central Devon magnetic interface (Fig. 3) does not necessarily require a fault for its interpretation but an extension of some of the arguments of Durrance (1985) would lead to the suggestion that this feature is due in part to thrust development, perhaps giving rise to an incipient extensional basin structure responsible further east for the Tiverton Trough (Fig. 3). It would be of interest to test the suggestions of Durrance (1985) elsewhere, particularly in the Devonian sequence of alternating and contrasting units of north Devon, where the lithological boundaries could coincide with strike faults. For example, a distinct Bouguer anomaly feature indicates a pronounced thickening of Permo-Triassic rock southwards across the Ash Priors Fault (Fig. 1), which could represent a reactivated fault between the Morte Slates/Pickwell Down Sandstones (Fig. 1). At the western end of this boundary the Bouguer anomaly data of Al-Sadi (1967 Fig. 3), suggest the dip may locally exceed the 40° anticipated from geological evidence. To the south, the increased Bouguer anomaly gradient G9 (Fig. 1) suggests the existence of faulting at depth east of the northern margin of the Tiverton Trough.

The relationship between the main geophysical lineaments in the Quantock Hills area and the postulated horizontal dextral displacement of 5km (Webby, 1966) along the Cothelstone Fault is interesting but complex. The Bouguer anomaly gradient zone G2 is displaced in the vicinity of the Cothelstone Fault but the feature appears to have little overall displacement away from this area (Fig. 1). Similarly the magnetic lineament M1-M3 (Fig. 1) is not apparently displaced significantly but if the two anomalies M2 and M3 are due to the same magnetic rock in the core of the North Devon/Quantock anticlines, a dextral displacement of about 9km is indicated. This last interpretation poses other problems in the understanding of the gravity gradient zone. These apparent inconsistencies would be explained if some of the postulated ESE-trending faulting occurred after the main phase of movement of the Cothelstone Fault.

The trend and magnitude of the lineaments suggest they

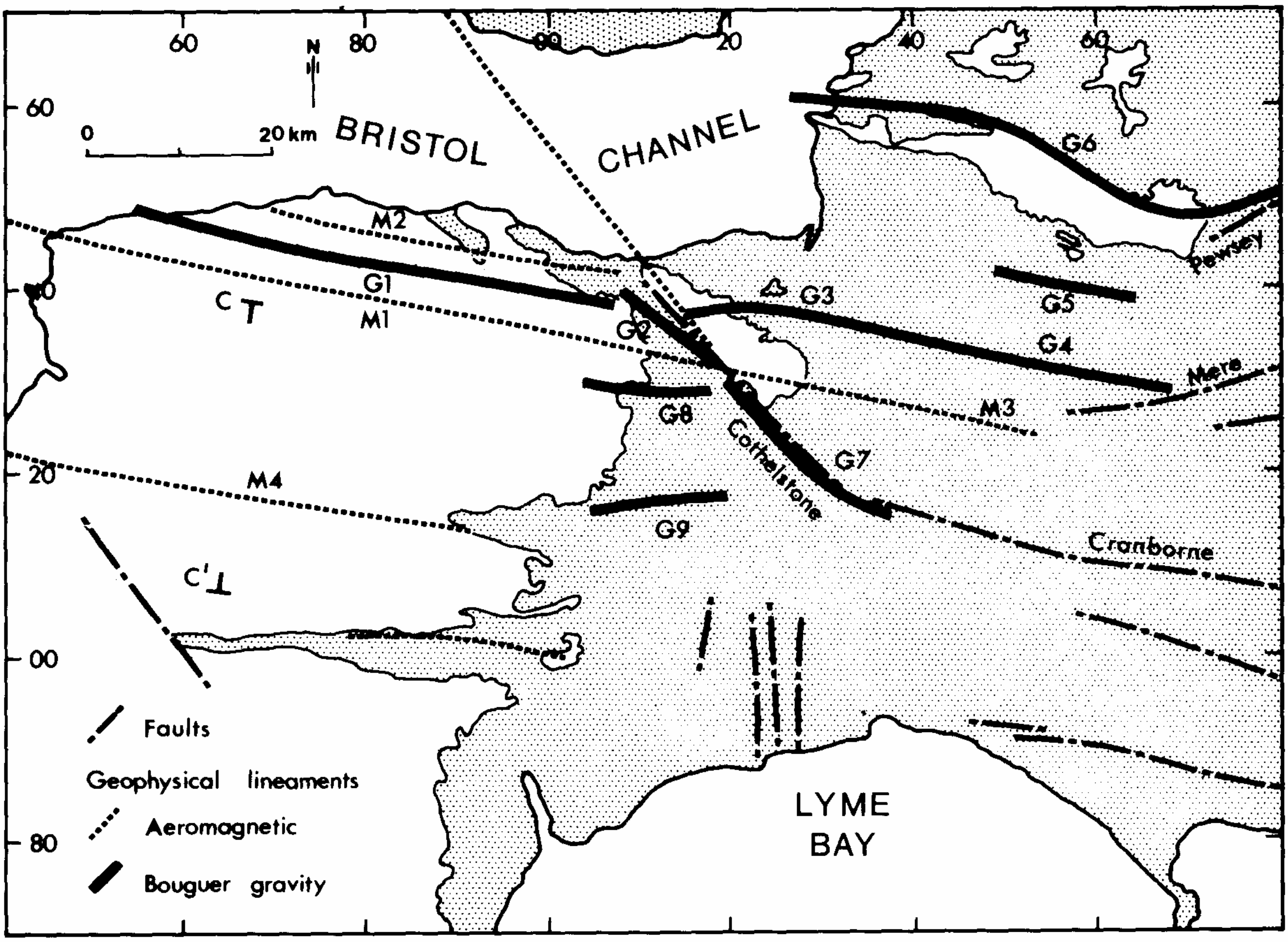

Figure 4. Map showing the geophysical lineaments discussed in the text and main faults affecting the Mesozoic sediments in the Wessex Basin.

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could be related to the basement faults which were reactivated during Mesozoic times in the Wessex Basin. Here Chadwick and others (1983) report that the major growth faults affecting Mesozoic sedimentation in the Wessex Basin (the Vale of Pewsey, Mere and Cranborne structures) can be traced from seismic reflection sections into the basement where they appear as southerly dipping thrusts or listric faults.

Conclusions

Regional Bouguer anomaly and aeromagnetic data for west Somerset and part of north Devon indicate several linear anomalies which are clearly related to the major geological structures in the area. The interpretation of the Bouguer anomaly data is complicated in part of the area by the range and similarity of the Mesozoic and Devonian rock densities. A southward extension of the Cothelstone Fault system is clearly indicated. In the Quantock Hills the anticlinal structure appears to be responsible for a significant deviation of the otherwise linear zone of Bouguer anomaly gradients (`the Exmoor gradient'). The ESE trend of this zone is repeated by at least two other lineaments defined largely by aeromagnetic data. This trend is parallel with that of the regional geological strike but the linearity and extent, both horizontally and vertically, of the geophysical features suggest the existence of strike faults perhaps reactivated at later dates. The faults could be the westward continuations of the basement structures which subsequently controlled the development of growth faults in the Wessex Basin.

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Read at the Annual Conference of the Ussher Society, January 1986

Soil gas surveying at Whitchurch Down, near Tavistock, Devon: an integrated approach to vein and fracture mapping

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Gregory, R.G. Durrance, E.M. and Mitchell, C.R. *1986. Soil gas* surveying at Whitchurch Down, near Tavistock, Devon: an integrated approach to vein and fracture mapping. *Proceedings of the Ussher Society, 6, 389-397.*

Vein-type Cu and Pb mineralisation occurs within faulted and thrusted Upper Devonian and Lower Carboniferous strata on the northern flank of Whitchurch Down, near Tavistock. Measurements of soil gas CO_2, O_2, Rn and ^{4}He have been taken on two traverses crossing major veins and faults.

CO_2 and O_2 measurements are clear indicators of structural and mineralogical control. Total Rn activity may be used as a guide to fracture-controlled gas migration, and when resolved into ^{220}Rn and ^{222}Rn activities, allows the classification of fractures into high or low permeability. Comparison of these results with the ^{4}He data allows further subdivision permitting the setting-up of a fourfold classification of fracture and overburden type.

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Introduction

Mineral exploration, using a few selected soil gases, is a well established technique. In particular, CO_2 and O_2 soil gas anomalies over oxidising sulphide ores are well documented (Ball *et at,* 1983; Lovell *et al.,* 1980). The signature of such occurrences is an increase in CO_2 together with a depletion in O_2, with respect to the atmospheric equilibrium concentrations of the gases. Depletion of 02 exceeds production of CO_2, so in simple cases there is a concomitant drop in the value of $CO_2 + O_2$ (Gregory and Durrance, 1985a). However, over non-mineralised faults and fractures, oxidation of CO, CH_4 and other species rising from anoxic environments in the crust also yields CO_2. Furthermore, variation in bacteriological activity can produce changes in the CO_2 and O_2 concentrations which may partly mimic these effects.

Other soil gases which have been used for mineral exploration include the radiogenic gases ^{220}Rn, ^{222}Rn, and ^{4}He. The main development of the ^{222}Rn surveying method has been in the search for hidden U orebodies, and the technique is well established. Gingrich (1984) provides an excellent review. However, most soil gas ^{222}Rn anomalies occur over fractures as these provide pathways for ^{222}Rn (half-life 3.825 days), originating at depth, to reach the surface. Enhancement of ^{222}Rn anomalies can be caused by localised precipitation of ^{238}U or ^{226}Ra from circulating groundwater in the fractures, or their overburden (Israel and Bjornsson, 1967). Soil gas samples over fractures can frequently yield high ^{220}Rn concentrations as well as 222Rn, indicating that the total Rn flux observed is a function of the background 238U and 232Th concentrations rather than solely due to U mineralisation. Moreover, with a half-life of only 55.6 secs, high levels of ^{220}Rn in particular indicate the presence of rapid transport and thus very permeable fractures.

Surveying techniques employing the measurement of 'He in soil gas also originated as a method of U exploration (Dyck, 1976; Reimer *et al.,* 1979). Recent studies by Butt and Gole (1984) and Gregory and Durrance (1985a, 1985b) have now shown that, as with Rn, the main factor controlling ^{4}He concentrations appears to be the presence of fractures. Consequently, the rationale that ^{4}He and ^{222}Rn are intimately associated with U mineralisation only applies where U deposits are located in, or intersected by, freely drained fractures.

If the gases are combined in a single integrated programme the different characteristics of CO_2, O_2, Rn and ^{4}He mean that more accurate location of mineralogical and structural features is possible. This approach was successfully tested by Gregory and Durrance (1985a, 1985b) utilising soil gas ^{4}He, CO_2 and O_2. In this study, the method has been extended to

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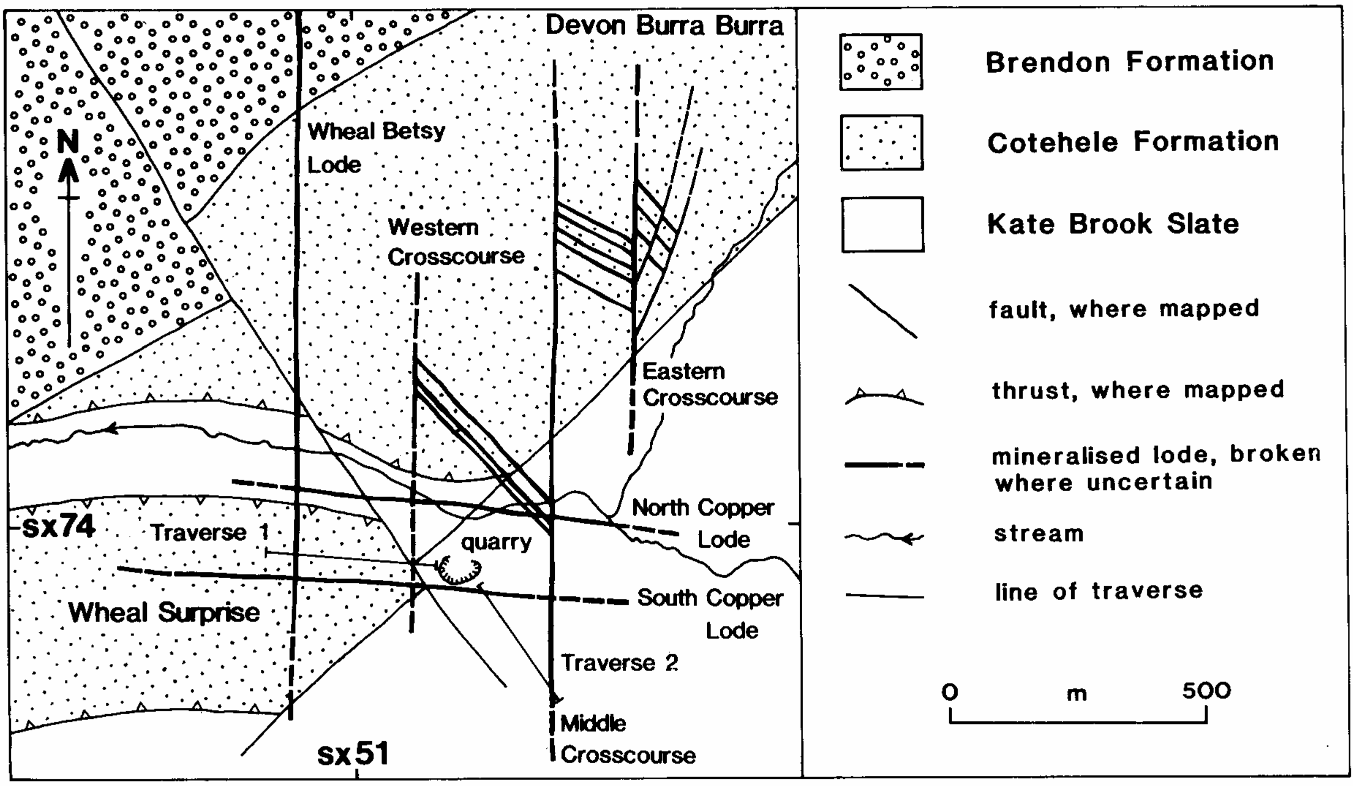

Figure 1. Geology and mineralisation at Whitchurch Down.

include Rn, using equipment that enables the identification of the two main isotopes, ^{220}Rn and ^{222}Rn.

Geological Setting

In the surveyed area, three major stratigraphic units are recognised (Bull, 1982). The southern part is composed of the Upper Devonian Kate Brook Slate Formation, but towards the northwest this is overthrust by the Lower Carboniferous Cotehele and Brendon Formations (Fig. 1).

Dines (1956) describes the location and distribution of known lodes from two mines in the area. Wheal Surprise, also known as Whitchurch Down Consols (SX 498 740), and Devon Burra Burra, or Gatepost Mine (SX 514 742) were both exploited for Cu.

Wheal Surprise, for which no production records are known, appears to have only been tried in a small way. A drainage adit and three shafts have proven two Cu lodes, known as North Copper and South Copper Lodes. The strike of both lodes is 092°, and North Copper Lode dips 72°S.

Devon Burra Burra shows greater development: two access shafts and underground workings exploited at least eight NW-SE trending Cu lodes (part of the early E-W trending set), and three N-S late mineralised crosscourses. The positions of the veins noted by Dines (1956) are shown on Figure 1.

A fourth crosscourse, west of the Devon Burra Burra workings, represents the southward continuation of the Wheal Betsy Lode (SX 510 812), which at Wheal Betsy dips 64°W. Geophysical surveying by Durrance (1973) confirmed that the lode continued further south. Quantitative analysis of the magnetic anomaly over the lode suggests that here the lode dips 74°E, while electrical resistivity measurements showed that the lode is associated with a non-mineralised fracture zone up to 100m wide.

Although the mineralisation of the crosscourses is typically Pb dominated, there are no records of Pb mineralisation at Whitchurch Down. The Cu ores here are also poorly documented, but typical lodes in this area bear chalcopyrite with arsenopyrite. There is no significant occurrence of U mineralisation within the survey area.

Sampling and Analysis

Two traverses for soil gas sampling were established on Whitchurch Down. The work had two main objectives. These were (1) to trace the Wheal Surprise and Devon Burra Burra lodes, and the southward continuation of the Wheal Betsy crosscourse, and (2) to investigate the effect of the structural features identified by Bull (1982) on the soil gas geochemistry. Traverse 1 lies west of the quarry in the Kate Brook Slate, running approximately E-W for 320m from SX 5083 7395 to SX 5115 7392, while

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traverse 2 was run on a NW-SE line for 260m from SX 5124 7390 to SX 5140 7367 (Fig. 1). The sampling interval was 20m.

Sample Collection

Shallow soil gas samples were obtained using a 0.75m stainless steel probe, fitted with a brass manifold, which enabled soil gas to be collected in sample syringes for laboratory ^{4}He analysis, or be pumped to on-site analysers for Rn, CO_2 and O_2 determination. The design and operational procedure were described by Gregory and Durrance (1985b).

At each site, once purged, the probe was sealed by an on-off valve on the manifold, and duplicate samples of soil gas were collected for ^{4}He analysis in 100cm^3 disposable plastic syringes via a septum attached to the sampling head. The syringe needles were capped with rubber stoppers to minimise mixing with atmospheric air. The sample half-life has been determined at approximately 7.5 days (Gregory, 1986). Consequently, all ^{4}He analyses were performed within 24 hours of collection, so that no significant changes would have occurred.

After the collection of the syringe samples, the on-site analysers were connected in turn to the probe, and the shut-off valve opened. 100cm^3 of soil gas was then pumped into the sample chamber of the Rn detection unit. On completion of Rn activity measurements, the combined CO_2/O_2 analyser was attached to the probe, 400cm^3 of soil gas flushed through the system, and finally, 100cm$^{3'}$ of soil gas was retained within the analyser for the subsequent determination of CO_2 and O_2.

Determination of helium

Samples were analysed for 4He using apparatus similar to that described by Reimer *et al,* (1979). This utilises a Dupont 120 SSA leak detector to which a constant pressure inlet system has been added. Each analysis was bracketed by a standard comprising water saturated laboratory air (WLA) as defined by Butt and Gole (1984). The value of WLA was calculated using samples of field-collected atmospheric air (FAA) which may be assumed to have a ^{4}He content of 5240ppb v/v (Gluekauf, 1946). The resulting 4He data are given as ΔHe/ppb-FAA, the disequilibrium from the atmospheric concentration. Limits of precision and sensitivity are better than ±10ppb. This is comparable with the best spectrometers, and therefore with regard to internal standards, and FAA, the replication of results is good. However, inter-spectrometer accuracy may be more variable, and while the relative background and peak values are comparable, the absolute values determined are not. Because there are few specialist spectrometers of this type, it is impractical to run duplicate samples on different machines to determine the overall accuracy. The design and layout of the ^{4}He spectrometer are discussed in detail by Gregory and Durrance (1985b).

Determination of the isotopes of radon

The determination of Rn in the field was accomplished with an EDA Electronics RD-200 Radon Detector, the use of which was first described by Dyck (1969). The small equipment consists of a smabattery-operated digital alpha scintillometer, in which a photomultiplier system has direct access to the sample chamber. This chamber is equipped with inlet and outlet ports to permit the free circulation of soil gas. To allow for instrument drift the frequent use of a standard long-lived alpha-emitting radionuclide was employed in combination with an amplifier gain potentiometer to maintain a check on the calibration. Raw counts were then multiplied by a time-dependent scaling factor to account for drift between determinations of the calibration standard. The analytical error approximates to ± 5% of the true count rate.

At each sample site, an open-ended ZnS (Ag) scintillation cell was inserted into the sample chamber and the background reading of the cell determined from a series of successive one minute counts. Once this had been completed, the soil probe was connected to the inlet port of the counting chamber, and 100cm^3 of soil gas passed through the cell with both inlet and outlet ports open. Both ports were then closed and three successive one minute counts were made. Finally the cell was removed and flushed of Rn and its daughters by vigorous aeration and left for about one hour before re-use.

Both Smith *et aL* (1976) and Morse (1976) have shown that sampling of Rn activity over short intervals of time can allow the separate identification of both ^{220}Rn and ^{222}Rn activities from a single sample. Smith *et al.* (1976) demonstrated that during the first three minutes of counting, the count rate drops due to the decay of 220Rn. Thereafter, as ^{218}Po activity continues to grow from the decay of ^{220}Rn, the count rate will rise to an equilibrium level which is attained 3-4 hours after sampling. Thus, depending on the relative contributions to the total Rn content by each isotope, the count rate may either rise or fall in the first three minutes of counting. The raw data obtained from the digital readout can therefore be processed to yield activities due to total An, ^{220}Rn, and ^{222}Rn. However, the cell background must first be deducted from the count rate of the three 1-minute counts obtained to give true count rates (c1, c2, and c3). Total Rn activity may then be calculated as the average of these count rates:

$$\text{total Rn} = (cl + c2 + c3)/3$$

Gregory (1986) has shown that increasing the number of one minute counts to five or ten has a negligible effect on the values so obtained.

Because of the impracticability of waiting for the equilibrium level of ^{222}Rn to be attained, Morse (1976) computed an algorithm to determine the ^{222}Rn activity from the total Rn count, accounting for ^{220}Rn decay and ^{218}Po ingrowth:

$$^{222}\text{Rn} = (0.87 \times c3) + (0.32 \times c2) - (0.34 \times cl)$$

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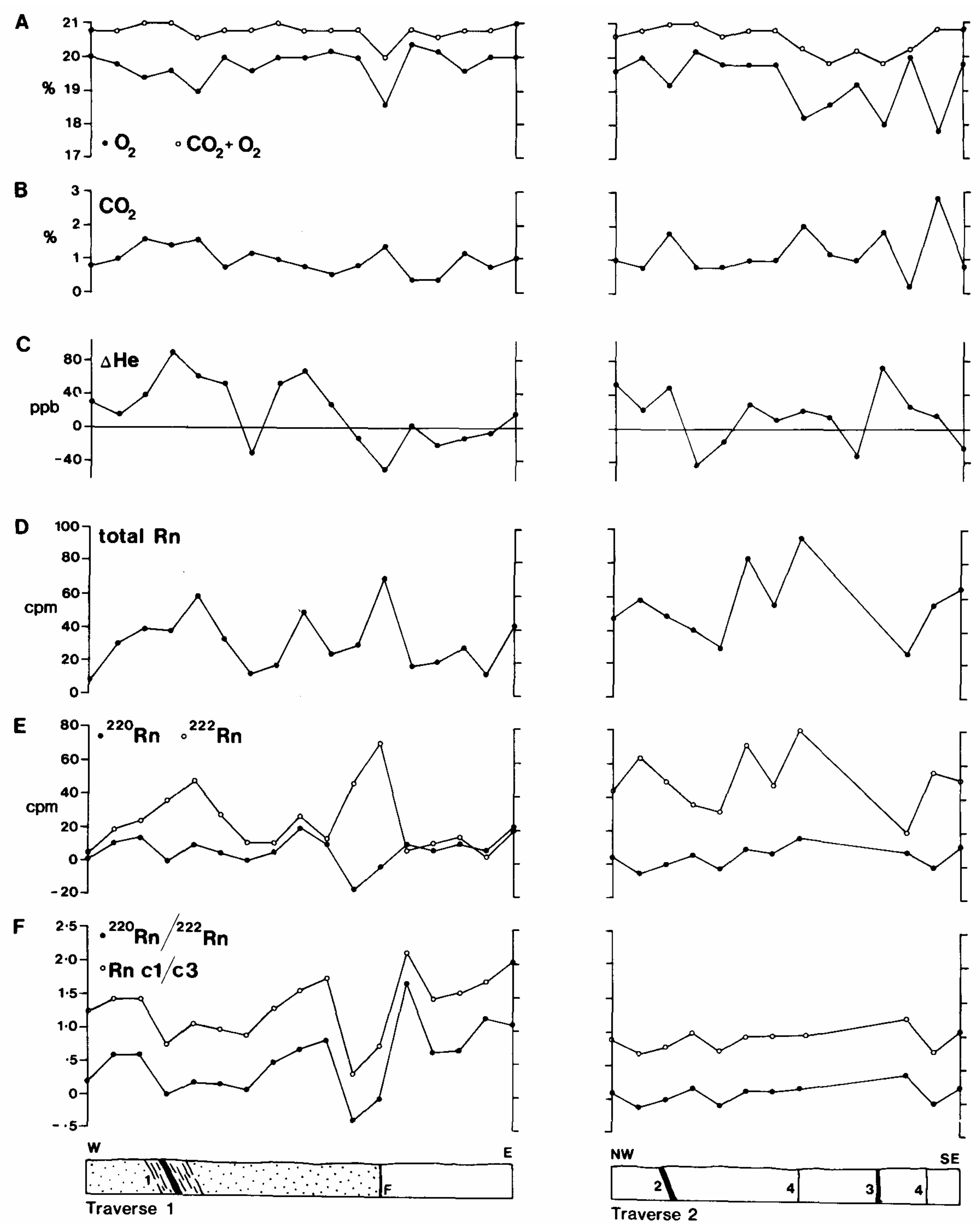

Figure 2. Soil gas measurements at Whitchurch Down.
A. O_2 and CO_2+O_2/%, B. CO_2/%, C. ^4He as ΔHe/ppb-FAA, D. Total Rn activity/cpm, E. ^{220}Rn and ^{222}Rn activities/cpm, F. Rn activity ratios ($^{220}Rn/^{222}Rn$ and Rn cl/c3).

Sections labelled: 1. Wheal Betsy Lode with associated fracture zone, 2. South Copper Lode, 3. Middle Crosscourse, 4. Veinlets associated with Middle Crosscourse, F. Fault.

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Consequently, the ^{220}Rn activity may then be derived: ^{220}Rn = total Rn - ^{222}Rn

Having evaluated the isotopic activites, determination of activity ratios can give information on the effective permeability of the gas source. A high $^{220}Rn/^{222}Rn$ activity ratio indicates high ground permeability, and is frequently associated with a high total Rn flux at structural discontinuities. However, $^{220}Rn/^{222}Rn$ activity ratios can be calculated in two ways. In the first, the calculated activities are employed, but the second method is simpler, employing the ratio of the first and third minute counts (Morse, 1976). The second method has the advantage that it eliminates the physically unrealistic negative $^{220}Rn/^{222}Rn$ ratios which occasionally arise in the first method. These are due to the calculation of a negative ^{220}Rn activity, which is simply a function of poor counting statistics in samples with low count rates. Both methods are presented here, and yield similar results.

Determination of carbon dioxide and oxygen

CO_2 and O_2 determinations were carried out in the field using a wet chemical method. The equipment was essentially an Orsat stack-gas analyser (Ball *et al,* 1983), which operates by absorbing each gas in turn. Replication of results was sufficiently good that no standards needed to be used. Full details of the modifications to the design of the Orsat gas analyser are given in Gregory and Durrance (1985b). The operating range for both gases is 0-40% ± 0.1% v/v. Variations in soil gas concentrations of these gases measure the disequilibrium from the atmospheric concentrations of 0.035% CO_2 and 20.95% O_2. Simple summation yields the value of $CO_2 + O_2$ at each sample site (Gregory and Durrance, 1985a, 1985b).

Results

Carbon dioxide and oxygen

It can be seen from Figure 2 that along Traverse 1, anomalously low O_2 values occur at sites 3-5 and 12, with correspondingly high CO_2 values at sites 3-5, 12 and 15. Only at site 12 was an anomalous value of $CO_2 + O_2$ detected. Along Traverse 2, low O_2 values occur at sites 3, 8, 11 and 13. These sites also exhibit anomalously high CO_2 in the soil gas measurements. Furthermore, a broad band of low $CO_2 + O_2$ values lies between sites 8 and 12. At all these anomalous sites, only on Traverse 1 at site 12 and sites 9-11 on Traverse 2 do the O_2 consumption/CO_2 production ratios come at all close to the theoretical values for the oxidation of sulphide species or anoxic gases. Thus while site 12 on Traverse 1 has a value of 1.7, sites 3-5 have an average value of only 1.1. Similarly, on Traverse 2, site 3 has a value of only 1.0, while sites 9-11 have an average value of 1.8.

Radon

Peaks in the total Rn count occur at sites 5, 9 and 12 on Traverse 1, and at sites 6 and 8 on Traverse 2 (Fig. 2). The peaks on Traverse 1 at sites 5 and 12 have high ^{222}Rn activity values coupled with low ^{220}Rn activities, while the peak at site 9 shows a near equal contribution from both isotopes. Another peak on Traverse 1 at site 11, is, however, owever, false. This arises simply as a function of poor counting statistics giving a large negative ^{220}Rn activity, and should be ignored. Overall, it should be noted that on Traverse 1 west of site 10, the ^{222}Rn activities are all higher than ^{220}Rn activities, while east of site 12 the values are approximately equal for each isotope, and at sites 13, 16 and 17, the ^{220}Rn activities are greater than the ^{222}Rn activities. This patterns is reflected in the $^{220}Rn/^{222}Rn$ and c1/c3 activity ratios. These ratios also show distinctly low values at sites 4 and 12, suggesting the presence of discrete regions of reduced permeability/diffusivity.

Patterns within the results from Traverse 2 are less easy to recognise because of three missing values at sites 9-11, due to problems with the cell chamber inlet valve. However, distinct peaks in total Rn occur at sites 6 and 8, with smaller peaks at sites 2 and 14. ^{220}Rn activity peaks occur at sites 2, 6 and 8. All ^{222}Rn activities are low, at less than 20cpm. Consequently, variations in the $^{220}Rn/^{222}Rn$ and cl/c3 activity ratios show no significant trends.

Helium

At first sight, the observed ΔHe values appear erratic (Fig. 2), but because of the nature of 4He in the pedological environment, both positive and negative A He anomalies are important, as are large discontinuities between adjacent values. Two broad positive ΔHe peaks occur on Traverse 1: at site 4, which lies within the broad zone of high CO_2 values, and at site 9 which otherwise only shows a Rn anomaly. Strongly negative A He values occur at sites 7 and 12. As with Rn, it appears possible to differentiate between the Δ He signature east of site 12, where A He values are close to zero (mean -6ppb), with the west of site 12, where Δ He values range up to 88ppb, with a mean of 35ppb. The difference between the mean values east and west of site 12 exceeds one standard deviation of the ΔHe results.

Along the length of Traverse 2 two large discontinuities in the ΔHe values are found at sites 3 and 11. These sites do not show Rn anomalies, but have anomalous CO_2 and O_2 values. In particular, site 11 shows a low value of $CO_2 + O_2$

Discussion

Traverse 1

Two parts of this traverse show anomalous values for all four of the gases measured. The first of these is a broad zone between sites 3 and 5. In contrast, the second is a sharp feature at site 12. Other less well defined anomalies also occur at site 9 (ΔHe and Rn), site 15 (CO_2 and O_2), and site 7 (ΔHe).

Explanation of the anomalies detected between site 3 and site 5 is straightforward: they clearly lie above the Wheal

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Betsy Lode and the associated fracture zone shown by Durrance (1973). The lack of supporting data from either the CO_2 + O_2 plot or the determination of the O_2 consumption/CO_2 production ratios suggest modification of the soil gas response by anoxic gases such as CO.

Coversely, the origin of the anomaly at site 12 appears unlikely to be related to mineralisation, as the ΔHe value is strongly negative. This ΔHe value suggests that the soil gas anomalies are simply related to a non-mineralised fracture. Certainly, negative ΔHe values elsewhere have been noted over fractures with a high overburden soil moisture content (Gregory and Durrance, 1985a; Gregory, 1986). This deduction is also supported by the sudden decrease in the ^{220}Rn/^{222}Rn and cl/c3 activity ratios which occur at this site: enhanced values of ^{222}Rn activity are usually associated with fractures, while negligible ^{220}Rn activities indicate restricted soil permeability, which may arise from soil moisture retention. Moreover, Durrance (1973) failed to detect the presence of further mineralisation east of the Wheal Betsy Lode. Therefore any projected southerly extension of the Devon Burra Burra Western crosscourse appears to be invalid. The $_{02}$ consumption/CO_2 production ratio at site 12 is 1.7, a value which also suggests oxidation of CO. However, caution is needed here because of the polymetallic nature of the ores: O_2 consumption/CO_2 production ratios for the common sulphide minerals typically lie between 1.875 and 3.0.

Support for the view that site 12 lies over a fault comes from the geological mapping of Bull (1982), and the overall distribution of the ΔHe and Rn data. ΔHe values west of this point are generally strongly positive in value, and ^{220}Rn activities are less than ^{222}Rn activity, while Δ He values to the east are close to zero, and $_{220}$Rn activities are seen to exceed those $^{\text{of}}$ ^{222}Rn. Both these observations suggest higher effective permeability of the soil east of site 12, with variation in the degree of soil aeration reflecting changes in the drainage character of the underlying rock type. The geophysical measurements made by Durrance (1973) gave no specific information about the presence of fractures.

Of the three sites which record poorly defined gas anomalies, site 9, which has a ^{220}Rn activity almost as large as the ^{222}Rn activity, and also shows a positive ΔHe value, may be a small non-mineralised fracture. The low O_2 value at site 15 does not appear to be particularly significant when compared with values recorded elsewhere on the traverse, so the status of site 15 as anomalous is equivocal. Finally, the negative ΔHe value found at site 7 probably reflects poor drainage of the soil at this locality.

Traverse 2

At sites 3, 8 and 11 three gases show anomalous values, while at site 13 only CO, and O_2 are anomalous. Although site 3 does not have an anomalous $CO_2 + O_2$ value, it otherwise possesses a signature which is typical of oxidising sulphide mineralisation, the ^{4}He peak taking the form of a positive He value superimposed on sharply dropping background values. The result is a large discontinuity in the plot of ΔHe at this site. A similar pattern of soil gas anomalies is seen at site 11, but this also lies within a belt of low $CO_2 + O_2$ values. Because of the good 'He responses and associated CO_2 and O_2 values, this site is also interpreted as lying over sulphide mineralisation. Site 3 lies on the eastward continuation of South Copper Lode and site 11 on Middle crosscourse. O_2 consumption/CO_2 production ratios at these sites are variable. Site 3 has a low value of 1.0, while sites 8, 11, and 13 have values of 1.4, 1.7, and 1.1 respectively. Ratios less than those previously described as occurring over oxidising sulphide mineralisation, or due to oxidation of reduced gases, suggest partial re-equilibration of O_2 at shallow depths. Although a peak in ^{222}Rn activity occurs 20m west of site 3, this value is due to a negative ^{220}Rn activity caused by poor counting statistics.

Because there are no records of veins which could act as a source for the broad $CO_2 + O_2$ anomaly between sites 8 and 12, and as the traverse lies at only a small angle to Middle crosscourse, it is possible that the anomalies found at sites 8 and 13 lie over mineralised veinlets sub-parallel to this vein. The only Rn anomaly within this set of data lies over site 8, and is due solely to ^{222}Rn activity. A similar peak in total Rn activity at site 6 has no CO_2 or O_2 anomalies, and only a minimal ΔHe peak. It is difficult to decide whether these sites lie over mineralised or non-mineralised fractures.

Conclusions

Anomalous CO_2 and O_2 soil gas concentrations occur at all the sites where veins or fractures have produced anomalous 'He and Rn values. The typical response in all these cases is a drop in the value of O_2 with a concomitant rise in CO_2. This is shown by the large negative Pearson Correlation between the two gases given in Table 1. Plots of CO_2 + O_2 for these sites are, however, only partly useful as some sites which otherwise show clear evidence of mineralisation do not demonstrate the drop in $CO_2 + O_2$ normally expected. We can only conclude that some physico-chemical parameters which have not been measured have affected the simple relation between these two gases. The most likely explanation from this area of vein-type mineralisation is the contribution to the soil gas anomalies of anoxic gases. Indeed, the corresponding O_2 consumption/CO_2 production ratios also show depressed values which suggest either the presence of CO in the soil gas or O_2 re-equilibration in the soil at shallow depth.

As in other areas, positive ΔHe values have been shown to occur in the soil gas over mineralised veins, or fractures with a low soil moisture content. Over fractures associated with a high soil moisture content, equilibration between groundwater and soil gas causes partial

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	CO_2	O_2	CO_2+O_2	ΔHe	Total Rn	^{220}Rn	^{222}Rn	^{220}Rn/^{222}Rn
O_2	-0.860 -0.873 -0.868							
CO_2+O_2	-0.090 -0.111 -0.182	0.586 0.582 0.646						
ΔHe	0.197 0.338 0.244	0.123 -0.397 -0.131	0.552 -0.246 0.116					
Total Rn	0.557 0.417 0.476	-0.716 -0.438 -0.586	-0.513 -0.133 -0.405	0.116 0.181 0.081				
^{220}Rn	-0.040 -0.038 -0.056	0.171 -0.117 0.074	0.271 -0.437 0.060	0.297 -0.103 0.200	0.138 0.556 0.157			
^{222}Rn	0.542 0.503 0.496	-0.755 -0.473 -0.611	-0.613 0.001 -0.422	-0.039 0.253 0.009	0.867 0.953 0.931	-0.374 0.280 -0.214		
^{220}Rn/^{222}Rn	-0.466 -0.365 -0.355	0.511 0.169 0.398	0.256 -0.500 0.206	-0.047 -0.098 -0.018	-0.264 0.042 -0.381	0.722 0.789 0.676	-0.609 -0.239 -0.625	
Rn cl/c3	-0.418 -0.397 -0.353	0.499 0.200 0.402	0.309 -0.497 0.222	0.033 -0.120 0.037	-0.255 0.050 -0.381	0.826 0.802 0.762	-0.655 -0.236 -0.658	0.963 0.997 0.971

Table 1. Pearson Correlation Matrix of Soil Gases for the Whitchurch Down Survey. For each correlation, three values are given. In descending sequence within each block of values, these are for Traverse 1 (17 sites), Traverse 2 (14 sites, 11 for Rn measurements), and all data (31 sites, 28 for Rn measurements).

depletion of 'He, leading to negative ΔHe anomalies. In addition, this study has shown that where a fault brings different rock types together, the measured ^{4}He values can be of contrasting nature either side of the contact. Generally, the association of ^{4}He with fractures leads to a positive Pearson Correlation with the similarly emanative gases CO_2 and Rn (Table 1). However, because of the complex behaviour of 'He and Rn, these values may be subdued with respect to CO_2. Nevertheless, it is clear that incorporating 'He measurements with integrated soil gas surveys is particularly useful in areas of variable soil moisture content.

Rn surveying has proved equally successful in identifying structural and mineralogical controls on soil gas behaviour. Total Rn activities are useful in providing a broad picture of fracture-controlled gas migration, but separate determination of ^{220}Rn and ^{222}Rn activities, together with the ^{220}Rn/^{222}Rn (or cl/c3) activity ratios gives results which can be used to distinguish between fractures with high or low permeability. They may also be directly compared with the results obtained by 'He surveying to classify fractures into four types, as functions of permeability and soil moisture content (Figure 3). Type 1 fractures may be classified as having high permeability and low soil moisture, and are therefore well-drained. Such a fracture may be interpreted from the results seen at site 9 on Traverse 1. Type 2 fractures exhibit higher soil moisture contents than Type 1, resulting in a negative ΔHe value, while Type 3 fractures are the reverse, with low permeability, but a low soil moisture content indicating good drainage above the fracture-overburden interface. A typical Type 3 response may be seen on Traverse 1 at site 4, where a drop in the

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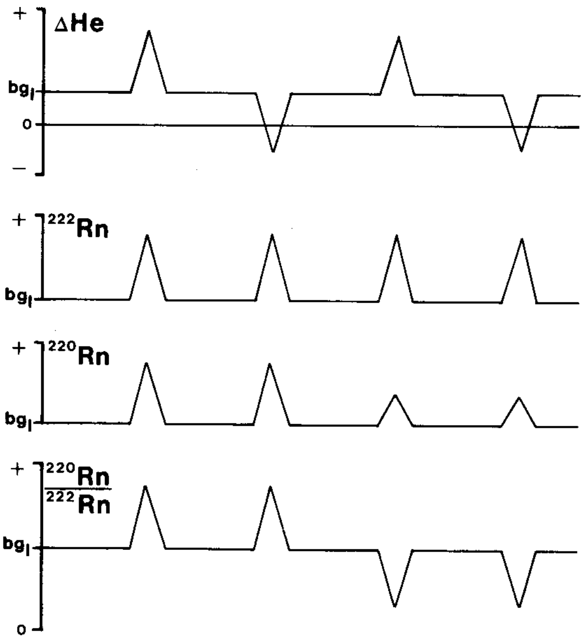
$^{220}Rn/^{222}Rn$ activity ratio is combined with a high A He value. The final permutation of these conditions, for fractures of Type 4, is of low permeability and high soil moisture. This is found at site 12 on Traverse 1. Fracture Types 1 and 4 are likely to be the most commonly encountered, as they represent the two extreme cases of soil and fracture conditions. Types 2 and 3 are modified responses due primarily to topologically controlled changes in the surfaces drainage pattern.

The use of a calculated ^{222}Rn activity, without calculating the ^{220}Rn activity is not recommended, as we have shown here that with low count rates, and consequently poor counting statistics, some ^{222}Rn activities may appear high, but can be associated with erroneously negative ^{220}Rn activities.

Although the relative amounts of ^{220}Rn and ^{222}Rn may be displayed either as calculated activity ratios or as the ratio of first minute and third minute counts, both methods yield similar results, as can be seen from Figure 2 and Table 1. The apparent advantage that the activity ratios show precisely where ^{220}Rn exceeds ^{222}Rn is countered by the fact that physically unrealistic negative values occur where negative ^{220}Rn activities have been



Fracture Type	1	2	3	4
Effective Permeability	high	high	low	low
Overburden Drainage	good	poor	good	poor

Figure 3. Classification of fractures by permeability and overburden soil moisture characteristics. All soil gas peaks are qualitative and relative to the local background (bg_1) for each fracture type.

calculated. There is thus no apparent reason to prefer the use of one method over the other.

Finally, it has been demonstrated that the determination of several soil gases is needed to permit detailed interpretation of areas of complex fracturing and mineralisation. The measurement of soil gases of different nature and behaviour can negate the problem inherent in simple gas surveying that many factors which control the response of the gas or gases measured are neither fully understood not easily quantifiable.

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Read at the Annual Conference of the Ussher Society, January 1986

Geochemical evolution of the Land's End granite (south-west England) in relation to its tin potential in the light of data from western marginal areas.

G. van MARCKE de LUMMEN

van Marcke de Lummen, G. 1986. Geochemical evolution of the Land's end granite (south-west England) in relation to its tin potential in the light of data from western marginal areas. *Proceedings of the Ussher Society, 6,* 398-404.

The Land's End granitic stock (Cornwall, south-west England) is composed of two main facies: *a* coarse grained megacrystic (locally poorly megacrystic) facies and a fine grained facies. Aplite and microgranite dykes cut across the granite itself and the country rock. A small cupola of medium grained, poorly megacrystic granite accompanied by a pegmatitic facies and microgranite dykes is found at Porthmeor Cove a few metres away from the main pluton. The evolution of major and trace elements shows the following crystallization sequence: (1) coarse grained granite, (2) Porthmeor cupola, (3) poorly megacrystic granite, and (4) fine grained granite and the microgranite dykes.

The granite has a coherent major and trace element chemistry typical of calc-alkaline (S-granite), shallow depth, highly fractionated igneous rock enriched in Rb, Ba, K, F, Sn... Trace element modelling has been used to test the efficiency of the fractional crystallization process during the magmatic differentiation. Interaction with late stage residual fluid has led to variation of Rb, Ba, K and Sr content and also to local K metasomatism and albitization. Whole rock analyses of 16 granites - including trace and rare earth elements - are given.

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Introduction

The Land's End granite is one of the six large granitic stocks exposed in Cornwall and Devon (south-west England). All these plutons are related to the buried Cornubian batholith which intruded 280-290 m.y. ago into a sequence of interbedded metapelites and meta-volcanic rocks of Devonian and Carboniferous age (Jackson *et al.,* 1982; Darbyshire and Shepherd, 1985). The emplacement of the batholith coincides with the formation of the Cornubian tin field from which considerable quantities of tin, copper, tungsten, lead, arsenic and zinc have been recovered for years.

This paper outlines the major geochemical characteristics of the Land's End granite in relation to its tin potential.

General geological relationships

Regional geology

According to Jackson (1976, 1979), the geological evolution of the Land's End area can be summarized as follows:

i) During the Middle Devonian (?), deposition of sediments and extrusion of alkali and tholeiitic basalts in a submarine environment took place, accompanied by local hydration and spilitic alteration of the basic lavas.

ii) Low grade metamorphism and deformation during the Hercynian Orogeny (Upper Devonian?).

iii) Emplacement of the Land's End pluton during the Lower Permian. The granite intrusion caused thermal metamorphism up to the hornblende hornfels facies.

iv) Formation of major fracture systems in and around the periphery of the pluton accompanied by Sn, Cu and As mineralisation.

Main rock types

The pluton consists mainly of medium to coarse grained granite with minor exposures of fine grained granite (Fig. 1). Dykes of leucogranitic and aplitic rocks cut across the granite as well as the country rocks. The pluton is roughly circular and has sharp contacts with enclosing rocks. Detailed petrography and field relationships are given in Exley and Stone (1964), Hawkes and Dangerfield (1978) and Mount (1985).

a) The coarse grained, megacrystic granite (type B after Exley and Stone, 1982) is characterized by large (1 to 5 cm long) K-feldspar crystals in a groundmass of quartz, plagioclase (An < 30%), K-feldspar, biotite, muscovite and accessory minerals such as apatite, zircon, tourmaline and magnetite.

The medium grained, poorly megacrystic granite (B', ibid) has the same mineralogy as the B granite but contains fewer K- feldspar megacrysts; it could

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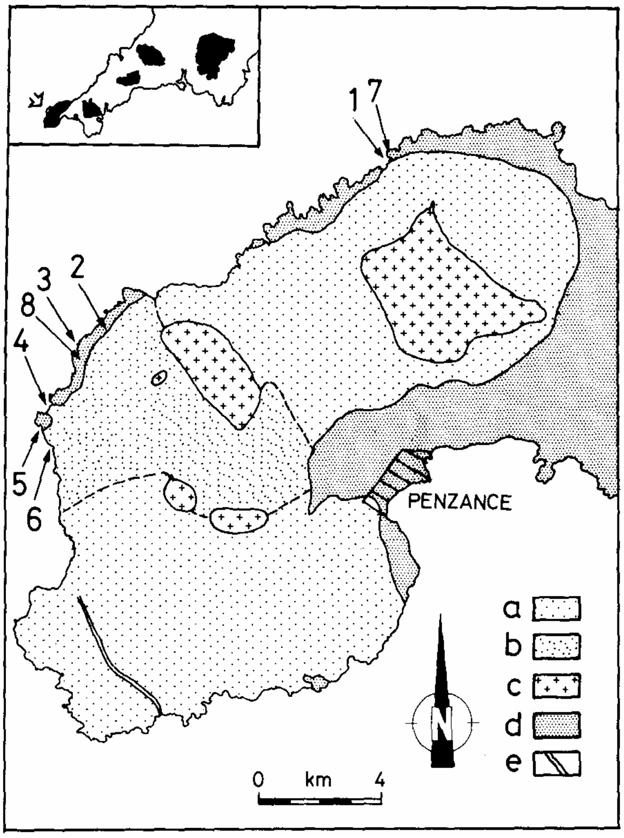

Figure 1. Schematic geological map of the Land's End granite. 1. Porthmeor Cove, 2. Geevor Mine, 3. The Crowns, 4. Porth Ledden, 5. Priest's Cove, 6. Porth Nanven, 7. Robin's Head, 8. Botallack Mine. a. coarse grained, megacrystic granite (B), b. medium grained, poorly megacrystic granite (B'), c. fine grained granite (C), d. pelite and metavolcanics, e. elvan. After Jackson (1976) and Hawkes and Dangerfield (1978). Inset: south-west granites in black.

a textural variation of B. It is well exposed in the central part of the pluton (Hawkes and Dangerfield, 1978).

c) The fine grained granite (C, ibid), characterized by the fineness of its grain, is much less abundant than the coarse grained granite (<10% of the pluton). The groundmass grain size is about 1mm. K-feldspar, plagioclase (An < 5%), quartz, biotite and muscovite are the major constituents. Apatite, zircon and fluorite are the common accessory minerals. Field relationships suggest that the tin mineralisation is closely related to this granite facies (Mount, 1985).

d) Leucogranite and aplite occur as dykes in the granite and in the country rock. The grain size is usually less than 0.5mm. They consist of K-feldspar, quartz, plagioclase (An < 5%) and biotite with apatite, magnetite and tourmaline. Topaz and amblygonite have also been reported (Exley and Stone, 1964).

e) The Porthmeor cupola A small (19 by 15m) biotite granite intrusion is exposed on the eastern side of the Porthmeor Cove. It is made up of megacrystic to poorly megacrystic granite. On the top of the pluton, there is a narrow pegmatitic body. A set of three dykes is associated with the pluton; they are numbered I, II and III following the chronology of their emplacement (Stone and Exley, 1984): (1) leucogranite dyke I, cut across by (2) granite dyke II which emanates directly from the pluton and (3) tourmaline microgranite dyke III which cuts across the two former.

Chemical composition of rocks

Major elements

Major elements show regular variations from coarse (B) to fine grained granite (C) (Fig. 2). Figure 2 shows that both the proportion of Al_2O_3 relative to that of FeO and MgO and the FeO/MgO ratio increase from granite B to granite C. SiO_2 increases whereas CaO MgO, Fe_2O_3, and TiO_2 decrease. Na_2O and K_2O are apparently constant. The K_2O content is high (5-6%) in granite B as well as in granite C and in microgranite dykes. Na_2O greatly increases in a few dykes where it reaches 5.5 to 6.5%. The evolution of the major elements reflects a magmatic differentiation in which type B granite represents the earliest rock and type C the most differentiated one. The latter granite composition is similar to that of most of the leucogranite and aplite dykes. Poorly megacrystic granites (type B') are set up in an intermediate composition (Fig. 2). The Porthmeor granite as well as dyke II are more differentiated than type B and the pegmatitic facies has a composition close to that of type

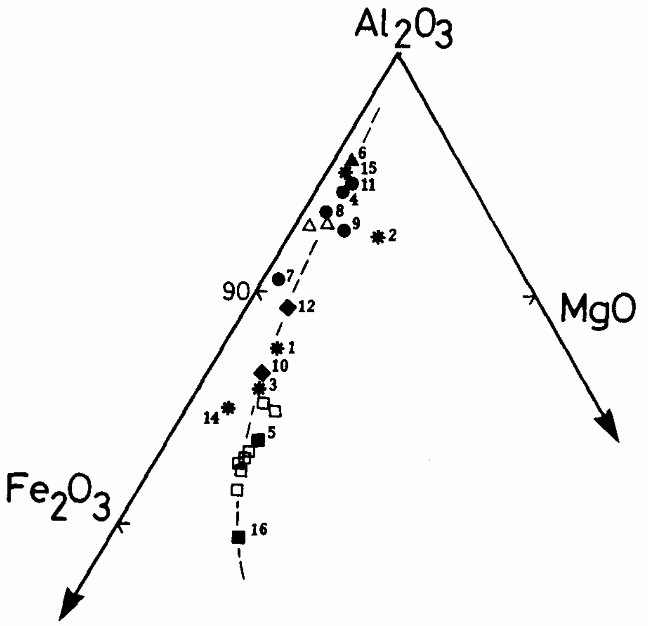

Figure 2. Al_2O_3: Fe_2O_3 (total iron): MgO (wt proportions) diagram showing the differentiation trend of the Land's End granite. Filled symbols: this study, open symbols: data from Wilson (1972) and Jackson (1976). squares: coarse grained granite, diamonds: medium grained, poorly megacrystic granite, stars: Porthmeor pluton (granite, pegmatite and dykes), circles: leucogranite and aplite dykes, triangles: fine grained granite. Dashed line: magmatic differentiation trend. Numbers refer to Table 1 and Appendix 1.

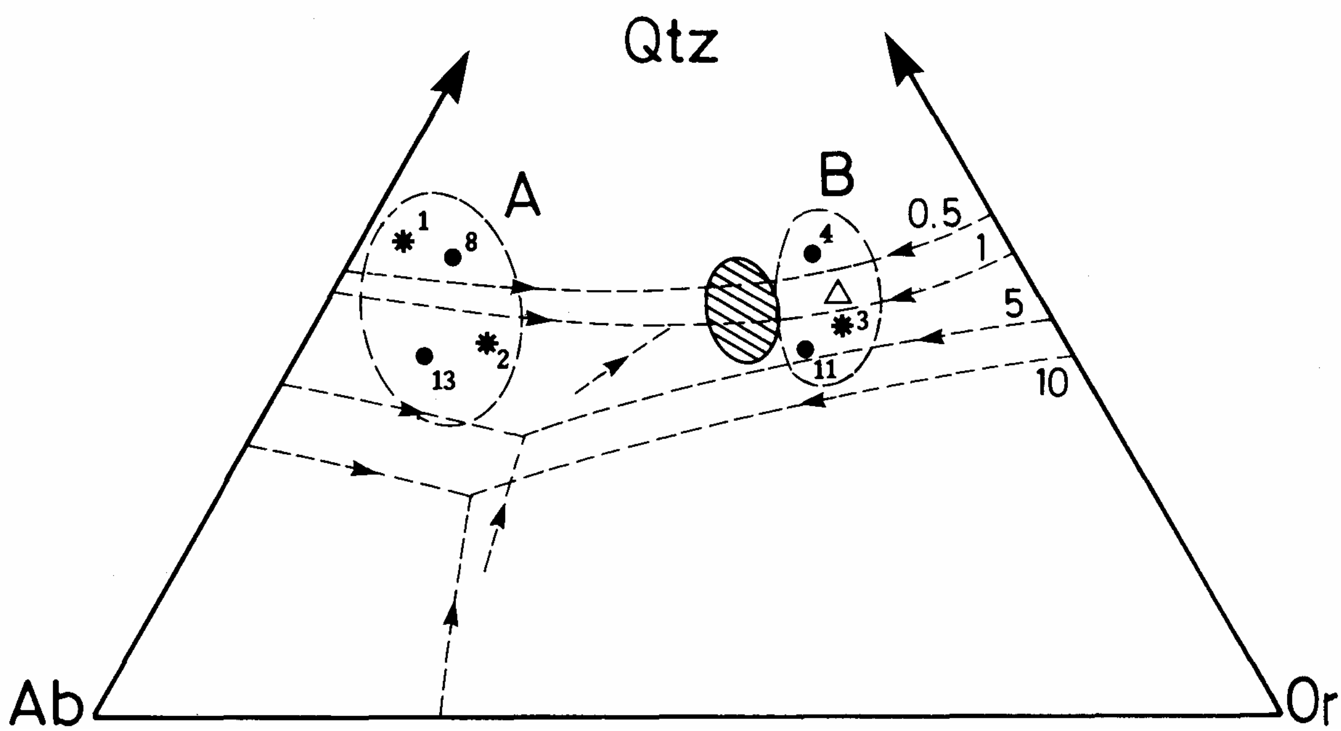

Figure 3. Quartz (Q) : albite (Ab) : orthoclase (Or) diagram. Dashed lines: cotectic lines and eutectic points (dry system) at 0.5,1,5 and 10kb. After Tuttle and Bowen (1958). Hatched area: coarse grained and fine grained granite composition field. Same legend at Fig. 2 (A) albitized and (B) K metasomatised rock.

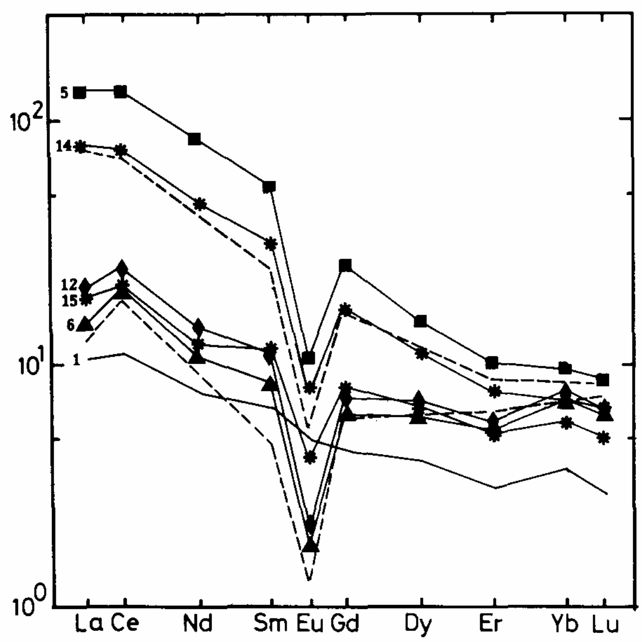

Figure 4. Coryell diagram showing the rare earth element pattern of granite samples. Same legend as Fig. 2. Solid line: example of albitized rock, dashed lines: calculated REE pattern of granitic rocks after Rayleigh law (fractional crystallization) for f=0.6 (40% crystallization) and f=0.2 (80% crystallization). The composition of the coarse grained granite (square) has been chosen as the parent melt composition. Its mineralogical (normative) composition is: quartz 38%, K-feldspar 32%, plagioclase 24%, biotite 6%, apatite 0.1% and allanite 0.01 %.

C. Most of the B and C granites contain more than 90% of normative quartz, plagioclase and K-feldspar. In the Q : Ab : Or diagram (Fig. 3) these granites lie close to the ternary minimum for water saturated magmas (between 1.0 and 0.5 kb.). However, a few samples, lying outside the Land's End granite composition field in Figure 3, are enriched in sodium and others in potassium. Their composition cannot be explained by simple magmatic differentiation and probably resulted from Na and K metasomatism.

Trace elements

The trace elements Zr, Y, Cr, Zn, V and the REE decrease with the magmatic differentiation (see Table 1). The fractionation of the rare earth elements decreases as well (Fig. 4): La/Yb = 21.7 and 3.2, La/Sm = 4.3 and 3.0 and Gd/Yb = 3.4 and 1.1 for granites B and C, respectively. The pattern of the heavy rare earth elements (HREE) is concave upward. A negative europium anomaly is exhibited and remains more or less constant. The Na-rich rocks do not display such a negative europium anomaly (Fig. 4).

The Sr and Ba contents decrease from granite B to granite C. Although boron and fluorine have not been analysed in the samples, the presence of tourmaline is indicative of relatively high boron content (probably up to 1% B) and the F and C I content in biotite shows that the granite was enriched in F relative to other Sn-W granites and that the fugacities of HF and HCl were high in the associated fluid (van Marcke de Lummen, 1985).

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Petrogenetic considerations

Magmatic differentiation

The variation of major elements shows a general magmatic differentiation trend in the Land's End granite. The crystallization was dominated by separation of plagioclase, K-feldspar and quartz as shown in Figure 3. The variations of MgO and TiO_2 can be accounted for by biotite separation which is the only Mg-Ti bearing phase. In the Porthmeor pluton, the main granite belongs to the general differentiation trend and is slightly more fractionated than granite B. After its separation from the Land's End magma, it evolved separately to give rise to the pegmatitic body.

Trace element modelling

The behaviour of trace elements such as the REE cannot be completely explained by the crystallization of major minerals. The decrease in the total REE indicates that the bulk partition coefficient for REE between the rock and the melt was greater than 1. This resulted from the crystallization of accessory REE-rich minerals. The concavity in the HREE pattern is due to apatite crystallization while the strong decrease in the LREE is accounted for by the formation of apatite and allanite and/or monazite. The effect of these minerals on the trace element contents can be deduced from calculated models. During fractional crystallization, the Raleigh law can be used to describe the REE concentration, C_0, in a melt relative to the concentration, Co, in the parent melt. C_1 is equal to $C_0\ f^{(D-1)}$ where f is the fraction of melt remaining and D the bulk distribution coefficient. We applied this fractionation law to test the validity of the fractional crystallization in the Land's end granite.

Because most of the granitic rocks of type B and C lie close to the ternary minimum in the Q : Ab : Or diagram, it can be assumed that their composition is not very different from that of the magma from which they

Table 1. Whole rock analyses of the Land's End granite

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
SiO_2	78.51	73.68	71.36	73.85	70.02	73.25	-	72.9	77.43	69.54	73.67	73.27	63.52	72.33	70.99	70.08
TiO_2	0.1	0.06	0.31	0.1	0.35	0.09	0.06	0.04	0.05	0.17	0.18	0.22	0.04	0.23	0.2	0.57
Al_2O_3	12.55	14.41	14.44	13.84	14.03	13.67	13.64	15.65	12.88	13.87	14.14	13.49	18.55	13.91	15.41	14.61
Fe_2O_3	1.56	0.73	2.08	0.74	2.28	0.6	1.4	1.03	0.81	1.98	0.69	1.47	3.53	2.28	0.76	3.03
MnO	0.02	0.03	0.06	0.02	0.04	0.01	0.02	0.02	0.02	0.03	0.04	0.04	0.07	0.03	0.04	0.07
MgO	0.29	0.53	0.33	0.14	0.57	0.1	0.09	0.15	0.26	0.3	0.16	0.21	0.49	0.25	0.11	0.84
CaO	0.38	0.86	0.58	0.5	0.94	0.52	0.36	0.59	0.51	0.39	0.64	0.36	0.68	0.54	0.52	1.09
Na_2O	5.75	6.1	2.49	2.1	3	3.1	1.94	5.52	5.28	1.78	2.87	3	6.44	2.43	3.28	1
K_2O	0.44	1.59	5.88	5.58	5.18	5.1	5.3	2.1	1.29	6.29	6.28	5.15	5.23	5.11	6.83	5.77
P_2O_5	0.13	0.36	0.19	0.1	0.23	0.08	0.18	0.3	0.14	0.33	0.3	0.19	0.15	0.08	0.23	0.37
L.I.	0.36	0.88	0.62	0.71	0.92	0.66	0.41	0.91	0.67	0.89	0.52	0.62	0.84	0.76	0.53	1.1
total	100.09	99.23	98.34	97.68	97.56	97.18		99.21	99.34	95.57	99.49	98.02	99.54	97.95	98.9	98.53
B	*			*		*	*		*	*	*			*		*
S	92	93	93	85	2	2	1	2	3	3	1	87	94	96	1	-
Cl	365	405	672	463	532	289	466	178	376	328	193	375	330	417	268	-
V	9	6	15	5	25	4	6	5	4	11	6	8	6	13	5	-
Cr	51	49	42	56	156	47	104	29	36	35	34	37	23	27	27	40
Ni	7	9	7	7	7	10	6	7	6	5	5	5	5	5	5	2
Zn	48	55	38	31	51	34	47	37	39	44	35	33	41	35	39	21
Rb	35	146	423	321	492	501	470	314	114	712	428	621	507	446	434	511
Sr	55	135	66	63	85	30	42	55	160	29	88	12	93	59	46	78
Y	6.88	12.66	17.08	-	24.76	11.8	-	6.86	-	-	-	14.72	6.5	18.97	12.42	-
Zr	34	40	74	38	122	25	30	15	52	57	28	24	30	73	28	139
Nb	22	10	22	24	21	23	19	79	4	25	23	25	71	22	23	-
Mo	9	10	9	10	10	10	10	10	9	9	9	9	9	9	9	n.d.
So	18	14	16	16	19	18	17	14	16	17	14	15	18	13	15	9
Ba	234	256	377	412	415	266	355	253	283	304	490	253	566	346	306	-
Ta	n.d.	n.d.	8	n.d.	9	9	10	7	5	n.d.	11	0	14	9	9	n.d.
W	14	17	7	9	9	10	6	15	3	12	6	15	4	9	3	3
Ph	n.d.	< 1	2	2	29	26	26	27	24	27	25	1	< 1	1	29	n.d.
La	3.3	8.07	22.14	-	41.66	4.6	-	3.12	-	-	-	6.47	1.33	25.1	6.01	-
Ce	9.28	17.28	53.43	-	102.49	16.23	-	8.66	-	-	-	19.83	9.04	59.24	16.69	-
Nd	4.27	8.67	23.44	-	47.62	6.31	-	4.11	-	-	-	8.29	2.71	26.6	7	-
Sm	1.24	2.03	5.32	-	9.78	1.56	-	1.41	-	-	-	2.05	0.84	5.82	2.12	-
Eu	0.35	0.62	0.56	-	0.76	0.13	-	0.44	-	-	-	0.15	0.25	0.53	0.29	-
Gd	1.14	1.9	4.02	-	6.61	1.59	-	1.2	-	-	-	1.88	0.87	4.36	2.05	-
Dy	1.22	1.9	3.12	-	4.56	1.84	-	1.39	-	-	-	2.14	0.96	3.37	2.02	-
Er	0.64	1.05	1.38	-	2.07	1.1	-	0.72	-	-	-	1.2	0.56	1.58	1.07	-
Yb	0.76	1.15	1.31	-	1.92	1.45	-	0.75	-	-	-	1.57	0.75	1.45	1.19	-
Lu	0.1	0.17	0.18	-	0.29	0.21	-	0.1	-	-	-	0.22	0.11	0.22	0.17	-

* boron present in tourmaline but not determined. n.d. not detected. - not determined. Total iron as Fe_2O_3. See appendix 1 for description and location. REE by DCP and other elements by XRF.

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crystallized. Thus, using the composition of granite B as that of the less fractionated magma and the partition coefficients given by Hanson (1978) and Cocherie (1985), the evolution of the REE during fractional crystallization can be computed (Fig. 4). The calculated REE compositions for f=0.6 fit quite well with the main granite of Porthmeor pluton. For f=0.2, they coincide well with the fine grained granite C as well as with the aplite dykes (not shown in Fig. 4). These rocks can thus be considered to have originated from the coarse grained granite by fractional crystallization. In the same way, the composition of the Porthmeor pegmatite can be calculated from the Porthmeor granite composition.

Late-stage residual fluids

The importance of late-stage fluids on granitic rocks of Cornwall has been demonstrated by Stone and Austin (1961), Exley and Stone (1964), Jackson (1979), and Stone and Exley (1985). These fluids were thought to be responsible for subsolidus growth of K-feldspar megacrysts through an early K-autometasomatism (Stone and Austin, 1961).

In the Q : Ab : Or diagram, most of the coarse grained megacrystic (B) and fine grained (C) granites lie close to the ternary minimum; but a few samples show a distinct K enrichment (Fig. 3). A difference is also found between the distribution of Ba, Sr and Rb (K/Rb) in the igneous rocks and the expected composition deduced from the fractional crystallization model (Fig. 5). The rocks are depleted in Sr and enriched in Ba relative to the theoretical composition deduced from granite B following the crystallization model. The evolution in the K/Rb ratio does not fit the model as well (Fig. 5). In the K metasomatised rocks, the K/Rb ratio is higher than in granite B. This ratio commonly decreases with increasing fractionation in granitic rocks (Taylor, 1965); and is usually less than 100 in tin granites whereas it is greater than 100 in barren granites (Tischendorf, 1977). The

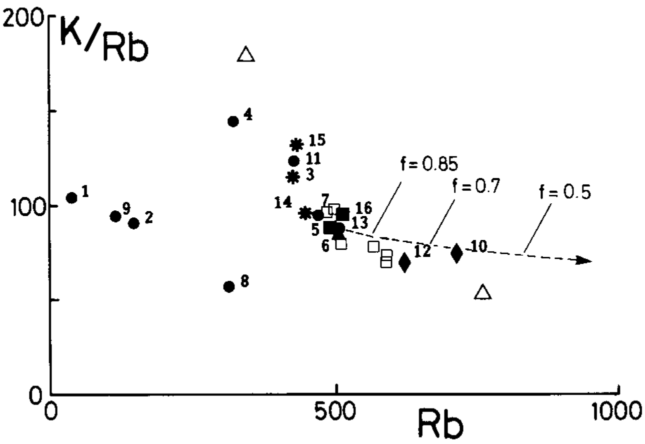

Figure 5. Plot of K/Rb ratio versus Rb (ppm). Dashed line: calculated magmatic differentiation trend. See also Fig. 2 and 4.

scattering of the evolved granite compositions plotted in Figure 5 - compared to those of granite B which are well grouped - can be interpreted as the result of a late-stage rock-fluid interaction. This fluid presumably was rich in K, Ba, and Rb (?) and depleted in Sr.

Albitization is another result of fluid circulation in the granitic rocks. However, this type of transformation seems to be restricted to the microgranite dykes (e.g. dykes I and III at Porthmeor Cove; see Fig. 3, Table 1 and Appendix 1).

Metallogeny of tin

Tin in granitic rocks could have been derived from different sources: (1) enrichment in tin of an initially unspecialized melt during fractional crystallization (magmatic differentiation theory, Lehmann, 1982), (2) melt inherits its geochemical specialization by partial melting of crustal rocks and magmatic differentiation causes only an enhancement of the initial anomaly (geochemical heritage theory, Lehmann, 1982) and, (3) tin was derived by hydrothermal leaching of upper crustal rocks (Jackson, 1979).

The specialized nature of the Land's End granite (enrichment in Rb, Li, B, F, Sn...) as of most of the tin-bearing granites (Stemprok, 1974; Imeokparia, 1984; Eugster, 1985) suggests a magmatic origin for its tin

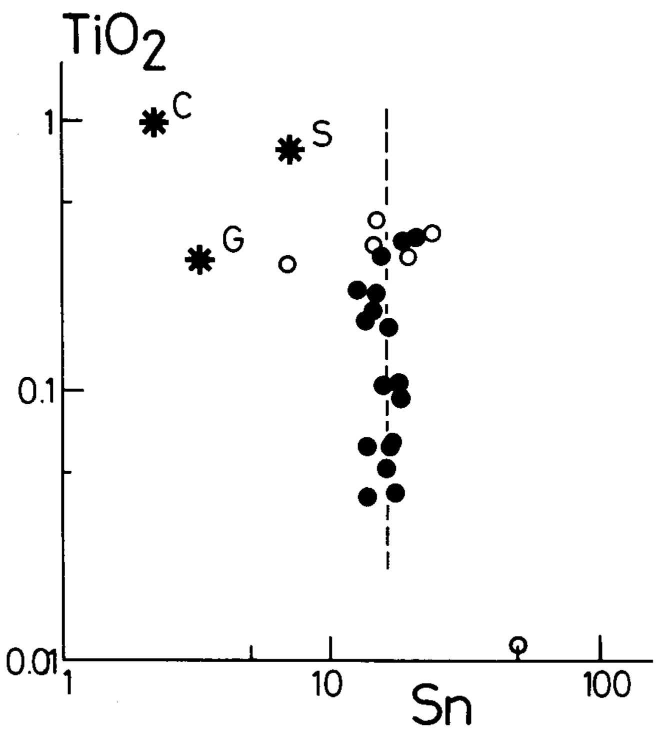

Figure 6. Bilog plot of Sn (ppm) versus TiO_2 (wt%). TiO_2 has been chosen as differentiation degree indicator. Circle: Land's End granite (filled circle: this study, open circle: analyses from Wilson, 1972 and Jackson, 1976). Dashed line: magmatic differentiation trend. Average composition for crust (C), granite (G) and schist (S) from Lehmann (1982).

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Appendix 1. Location and description of samples.

n°	location	description	mineralogy
1	Porthmeor Cove	dyke III*	qtz, ab, to, ace.
2	Porthmeor Cove	dyke I*	qtz, ab, acc.
3	Porthmeor Cove	dyke II	qtz, Ksp, pl, bi, mu, acc.
4	Robin's Head	microgranite dyke	qtz, Ksp, pl, bi, mu, to, acc.
5	Geevor Mine (level 15) Whisky central lode	coarse grained granite	qtz, Ksp, pl, bi, mu, to, acc.
6	Geevor Mine (level 15)	fine grained granite	qtz, Ksp, pl, mu, flu, to, acc.
7	Geevor Mine (level 16)	granite dyke	qtz, Ksp, pl, to, acc.
8	Porth Ledden	aplite dyke*	qtz, Ksp, pl, mu, acc.
9	Porth Ledden	granite dyke* ± pegmatitic	qtz, Ksp, pl, to, acc.
10	Priest's Cove	medium grained granite**	qtz, Ksp, pl, bi, mu, to, acc.
11	Priest's Cove	microgranite dyke**	qtz, Ksp, pl, mu, to, acc.
12	Porth Nanven	medium to coarse grained granite	qtz, Ksp, pl, bi, acc.
13	The Crowns	aplite dyke*	(qtz), Ksp, pl, bi(chl), acc.
14	Porthmeor Cove	medium grained granite	qtz, Ksp, pl, bi, mu, to, acc.
15	Porthmeor Cove	pegmatite**	qtz, Ksp, pl, bi, acc.
16	Botallack Mine (dumps)	coarse grained granite	qtz, Ksp, pl, bi, mu, to, acc.

qtz: quartz, Ksp:K-feldspar, pl:plagioclase, bi:biotite, mu:muscovite, to:tourmaline, chl:chlorite, flu:fluorite, acc:accessory minerals (magnetite, apatite, zircon...). See Table 1 for composition. * Albitized and ** K metasomatised rocks

mineralisation. However, there is no evidence for an increase in tin accompanying differentiation from an unspecialized material. The tin content of the Land's End granite is relatively constant (15 - 25 ppm Sn) from granite B to granite C (Fig. 6). The high initial 87Sr/^{16}Sr ratio of the Land's End granite (0.7133, Darbyshire and Shepherd, 1985) and its high d ^{8}0 values $^{(6180>}$ +11.7, Sheppard, 1977) clearly demonstrate a major involvement of upper crustal material. The high *d18O* value is compatible with derivation from pelitic rocks at depth (Taylor, 1978). The high level of large-ion lithophile (LIL) elements such as K, Rb, LREE... and low content of high field strength (HFS) elements (Nb, Ta, Y, Zr ...) are also consistent with crustal rather than mantle derived origin (Brown *et al.,* 1984). The magma which gave rise to the Land's End granite could, therefore, have been originated from partial melting of upper crustal material previously enriched in tin (geochemical heritage theory).

Conclusions

The geochemical study of the Land's End granite shows that it represents a highly fractionated rock of calcalkaline composition (S-granite) which crystallized from a residual melt rich in incompatible elements. The coarse grained granite is the less evolved term of the magmatic differention trend whereas the fine grained granite and the leucogranite and aplite dykes are the most fractionated. The differentiation proceeded by fractional crystallization of feldspar, quartz and biotite. The tin deposits associated with the granite owe their origin to magmatic processes. The melt probably inherited its specialization by partial melting of upper crustal rocks initially enriched in tin.

Interaction with late-stage fluids produced variations in the K, Ba, Rb, and Sr contents in the granite as well as local albitization and K metasomatism.

Acknowledgements The author thanks K. Gilbert, Managing Director and Mr C. Smith, Chief Geologist of Geevor Mine plc and Mr K. Beer of the British Geological Survey (Exeter) for geological advice and help during field work. Ph Sonnet, G. Power and an anonymous referee are gratefully acknowledged for their helpful comments. This study was financially supported by E.E.C. (Contract No MSM-127-B) and SPPS (Contract No MP/CE/13).

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Read at the Annual Conference of the Ussher Society, January 1986

The geochemistry of the foliated granitic rocks of Alderney, Channel Islands

T.S. BREWER G.M. POWER

Brewer, T.S. and Power, G.M. 1986. The geochemistry of the foliated granitic rocks of Alderney, Channel Islands. *Proceedings of the Ussher Society, 6, 405-411*

The foliated granitic rocks of western Alderney are shown to be talc-alkaline with Na_2O exceeding K_2O and most are peraluminous. Three rock units are defined: 1. the main quartz diorite showing little range of chemical composition; 2. the K-feldspar granite which forms patches within the quartz diorite and could be either the result of local fractionation or metasomatism; and 3. the Fort Tourgis type which has a distinctive chemical composition, particularly in terms of rare earth and transition elements and is likely to have resulted from contamination of the quartz diorite. These three units are interpreted as forming a single pluton.

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Introduction

The Channel Islands make up an important part of the Armorican Massif. They, together with nearby La Hague on the French mainland, probably contain the best geological evidence of the earlier Precambrian history of the Massif. However the evidence is fragmentary and in addition much of it still remains to be systematically described in a published form. This is particularly true for Alderney, the most northerly of the Channel islands.

The geology of Alderney may be divided into three main units:

1. The older foliated granitic rocks;
2. The younger diorites and granodiorites;
3. The Alderney Sandstones.

The Alderney Sandstones have been correlated with the Cambrian sandstones of La Hague (see Sutton and Watson 1970) and they rest unconformably on the foliated granitic rocks of unit 1. All three units are cut by a variety of dykes that range up to Carboniferous in age.

In this paper we report the results of a systematic petrographic and geochemical study of the Precambrian foliated granitic rocks of Alderney as a basis for comparisons with similar rocks elsewhere in the Armorican Massif.

Field relations

The foliated granites are the oldest rocks exposed on Alderney but there is no good evidence of their precise age. The only evidence of older rocks is as enclaves within the granites and exposed contacts with the younger unfoliated rocks of unit 2 are faulted. It should perhaps be noted here that the only published isotopic age determinations for these rocks (Adams 1976) are of an extreme reconnaissance nature and could be more misleading than useful. The foliated granitic rocks comprise most of the west of Alderney (Fig. 1) and are predominantly of quartz dioritic composition. Their main minerals are plagioclase, quartz and biotite in varying proportions. Hornblende is sometimes present and chlorite occurs as an alteration product. The foliation is defined by the shape orientation of aggregates of quartz and also of mafic minerals. Dark enclaves which are a characteristic feature of the quartz diorites may show considerable elongation parallel to the foliation. There is little evidence to support a primary igneous origin for the foliation. It is predominantly the product of deformation.

On the north coast a tidal platform provides good exposures. Here the quartz diorites are cut by a series of foliated aplitic veins and one NW-SE trending aplitic body reaching 100 metres in width. The aplites are cut in turn by unfoliated quartz porphyry dykes from one to fifteen metres in width. The quartz porphyries are not found to intrude the unfoliated diorites and granodiorite of unit 2. The south coast, in contrast, is made up of steep cliffs and access is limited. Again aplitic veins are a common feature and it is along this coast that the principal lithological variation from quartz diorite is found. This is a granitic rock which is distinguished by the presence of prominent euhedral K-feldspar megacrysts. There are no sharp contacts with the main quartz diorite and the K-feldspar granite shows a patchy development merging imperceptibly into quartz diorite. There appears to be some spatial association of K-feldspar megacrysts with aplitic veins but this is certainly not always obvious and may be unrelated to the occurrence of K-feldspar megacrysts elsewhere. Close to

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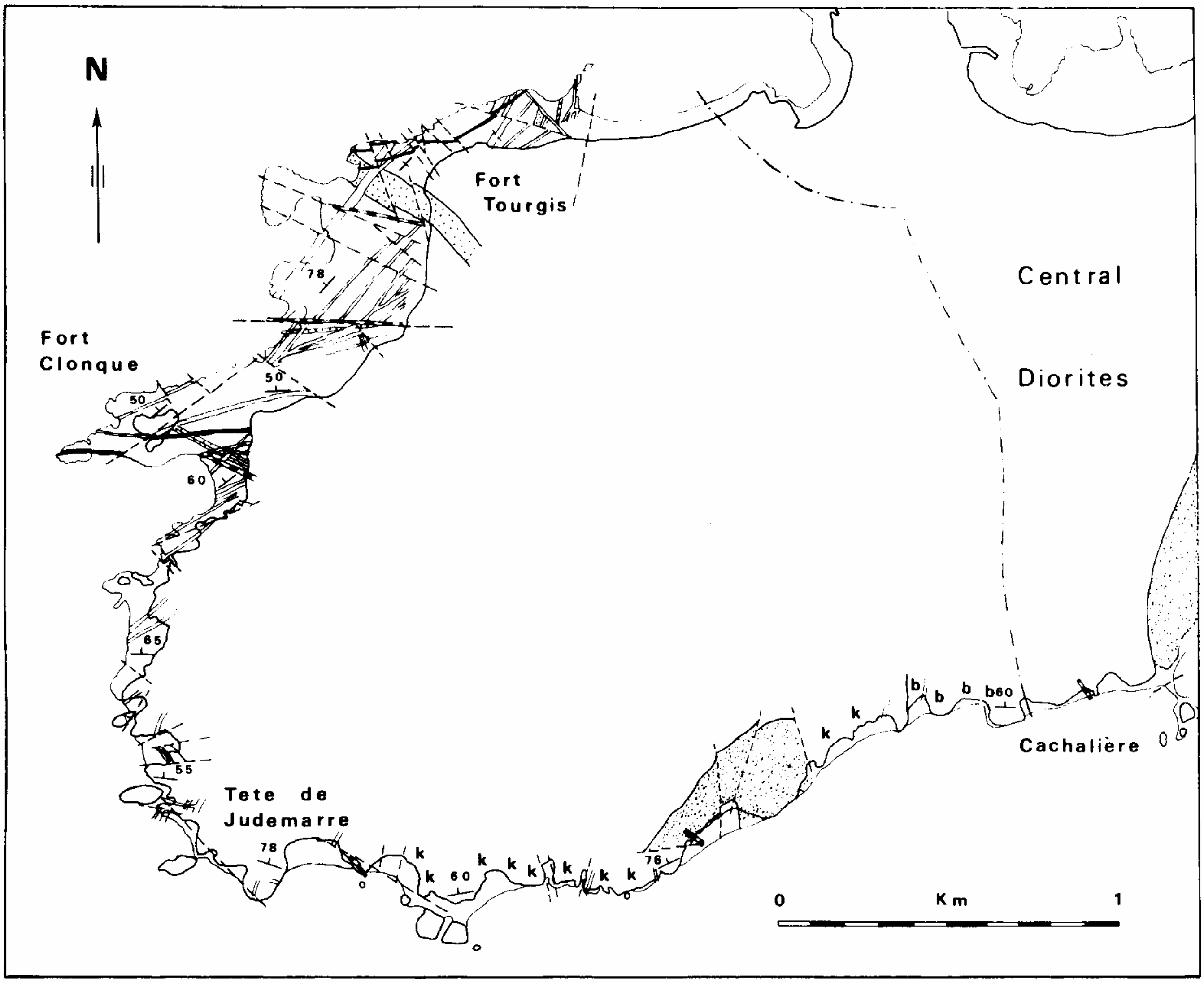

Figure 1. Geological sketch map of western Alderney. Aplites: stippled (many small veins omitted); Quartz porphyry dykes: blank; dolerites: black; lamprophyres: broken black; Alderney Sandstone: varied stipple; k: intermittent K-feldspar granite; b: biotite granodiorite; dashed line: faults; dash-dot lines: inferred faulted boundary of foliated quartz diorite and Central Diorites; symbol: strike and dip of foliation.

the faulted contact with the diorites at Cachaliere, at the centre of the south coast, a granodioritic body four hundred metres wide may be distinguished. This biotite granodiorite, contains no aplite veins but is cut by the quartz porphyry dykes. It may be a slightly younger intrusion than the other principal rock types. The intensity of development of the foliation in the granitic rocks of western Alderney is variable. In general the foliation is more intensely developed along the south coast. Two areas of lower strain may be recognised, one at Fort Clonque and the other at Tete de Judemarre (Fig. 1). The foliation varies in strike from about E-W to NE-SW and its present dip ranges from 58°-80° to the north. This may have been even steeper if tilting occurred before deposition of the Alderney Sandstones which now rest at an angle of about 20° on the foliated granitic rocks. The foliation is cut in places by centimetre wide steep, N-S trending shear zones showing brittle deformation. The whole of western Alderney is divided up into a series of fault bounded blocks. This is well displayed by the displacement of dykes on the tidal platform of the north coast, (Fig. 1).

Petrographic and geochemical studies have also highlighted a further lithological division of the foliated granitic rocks which is not particularly obvious in the field. It occurs from the north-eastern edge of the complex, around Fort Tourgis, at least to the fault 200m SW of the major aplite body (Fig. 1) and will be referred to as the Fort Tourgis type. In this area there is some evidence for shadowy, apparently partly assimilated dark enclaves within the quartz diorite.

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Petrography

In the succeeding sections dealing with the petrography and geochemistry of the foliated granitic rocks they will be considered in three principal divisions:

1. The main quartz diorites;
2. The K-feldspar granites;
3. The Fort Tourgis type.

However mention of the biotite granodiorite will also be made where appropriate.

It must be emphasised that in addition to the variable deformation of the rocks they also exhibit variable metamorphic modifications and both of these are super-imposed on the earlier lithological variations.

The main quartz diorites

The least deformed and altered rocks are medium to coarse grained and contain hornblende as well as quartz, plagioclase and biotite. The quartz occurs in equidi-mensional pool-like aggregates of grains with simple grain boundaries. Plagioclase is often euhedral and shows some evidence of zoning. It is always at least partly altered to sericite and is now always albite to sodic oligo-clase (An7-13) in composition. Hornblende may be subhedral but is often partly replaced by biotite. Biotite is the main mafic mineral and occurs both as large flakes and also as much smaller grains growing along grain boundaries. Some samples contain minor amounts of K-feldspar and accessory minerals include magnetite, sphene and quite frequently minor but important amounts of allanite.

All gradations from these rocks to more highly strained rocks may be observed. The main effects of increasing deformation appear to be the elongation of the quartz pools together with a decrease in the size of the quartz grains within them and the development of compex sutured boundaries between the grains. Plagioclase grains show microfracturing and possibly some rotation rather than recrystallisation. Accompanying these changes the mafic minerals also form more elongate aggregates mainly by the growth of biotite and the foliation is principally defined by the shape of the quartz strings and the mafic mineral aggregates.

The most prominent indication of metamorphism in the quartz diorites is the alteration of the plagioclase and hornblende. Plagioclase may become very cloudy and replaced by a sericitic white mica. Hornblende shows all stages of replacement by biotite and often only obvious biotite pseudomorphs after hornblende remain. These may also contain yellow epidote, magnetite and sphene. It is not clear if any of the biotite at all is of primary igneous origin. Less frequently some rocks contain chlorite as the replacive mafic phase with chlorite pseudomorphs after hornblende. These rocks contain no biotite developed along grain boundaries. Rarely both chlorite and large flakes of biotite may be found together in the same rock. There seems to be no regular geographic distribution of the chlorite-bearing rocks but it does appear that the more strongly deformed rocks are more likely to contain chlorite as the alteration phase.

K-feldspar granites

These rocks are much more leucocratic than the quartz diorites and only contain relatively minor amounts of mafic minerals usually chlorite possibly after hornblende. K-feldspar occurs as euhedral megacrysts of microcline perthite commonly up to 20mm in size and with irregular boundaries enclosing the other mineral phases of the rock. The effects of deformation and metamorphism on the granites are similar to those on the quartz diorites.

Fort Tourgis type

Rocks from the Fort Tourgis area contain the highest proportions of mafic minerals and hornblende is more commonly present. Hornblende occurs both as isolated euhedral grains and also as aggregates. The amphiboles within these clots often show regular polygonal grain boundaries suggestive of recrystallisation and contain regularly arranged blebs of magnetite within them. It is suggested that the majority of the amphibole may be of xenocrystic or xenolithic origin. Despite the more mafic nature of the rocks some do contain minor amounts of K-feldspar. Plagioclase is often very cloudy and full of finely divided opaque material. Some biotite does occur usually along grain boundaries but the more usual replacive phase is chlorite.

Table 1

	AW4	AW6	AW3	AW21	AW22	AW32
SiO_2	51.42	58.56	63.98	62.68	68.44	68.96
Al_2O_5	18.01	18.16	17.36	17.44	16.2	15.72
Fe_2O_3	3.37	2.64	1.84	1.81	1.44	1.51
FeO	6.44	3.69	2.85	3.49	1.88	1.93
MgO	5.88	4.03	2.55	2.5	2.19	1.8
CaO	7.23	5.18	3.15	4.4	1.01	2.02
Na_2O	3.47	4.13	3.78	3.84	4.24	3.7
K_2O	2.5	2.35	3.44	2.65	3.85	3.41
TiO_2	1.17	0.87	0.75	0.84	0.55	0.52
MnO	0.13	0.1	0.06	0.07	0.04	0.04
P_2O_5	0.38	0.29	0.24	0.28	0.15	0.15
Ba	469	777	900	923	704	859
Ce	65	35	62	79	53	48
Cr	206	86	37	40	37	27
La	10	4	30	39	19	21
Nb	10	9	8	9	9	8
Nd	48	25	28	31	20	18
Ni	55	27	10	10	7	6
Rb	86	98	106	91	124	102
Sc	18	6	9	1	8	8
Sr	633	708	691	741	417	543
V	129	98	58	63	55	47
Y	26	17	9	8	6	6
Zn	191	114	84	99	66	61
Zr	199	194	191	186	125	136

Major element analyses as weight percent recalculated to 100% water free. Trace elements in parts per million. AW4 xenolith; AW6 Fort Tourgis type; AW3 main quartz diorite; AW21 chlorite overprinted main quartz diorite; AW22 biotite granodiorite; AW32 K-feldspar granite.

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Biotite granodiorite

The biotite granodiorite is a medium grained rock with obvious biotite present in hand specimen. In thin section the biotite is seen to occur in euhedral areas possibly pseudomorphous after hornblende although no amphibole is now present. The main mineral is plagioclase (An 13) with quartz and subordinate K-feldspar, these last two minerals sometimes occurring as a graphic intergrowth between the plagioclase grains. No accessory sphene or allanite present.

Geochemistry

This work is based on thirty samples which have been analysed by X-ray fluorescence spectroscopy at Portsmouth Polytechnic for 10 major and 14 trace elements after the methods of Brown *et al.* 1973. Rare earth element analyses for eight samples were obtained by inductively coupled plasma emission spectroscopy (Walsh *et al.* 1981). A complete list of analyses may be obtained from the authors on request but representative analyses are given in Table 1. On the chemical plots that follow each of the four lithological groups discussed above are shown by a different symbol and in addition, six specimens that come from the main quartz diorite and that have well developed chlorite are given a distinguishing symbol.

The rock suite as a whole shows calc-alkaline affinities with sodium oxide in excess of potassium oxide. On a plot of A/CNK (molar Al_2O_3/ $(CaO+Na_2O+K_2O)$) against SiO_2 (Fig. 2) the majority of the samples would be

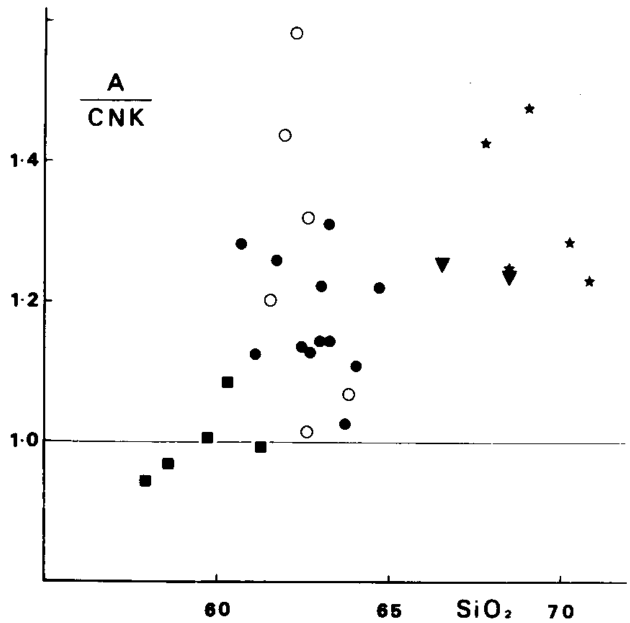

Figure 2. A/CNK (molar Al_2O_3/ $(CaO+Na_2O+K_2O)$) V. SiO_2 wt.% Symbols: square: Fort Tourgis type; filled circle: main quartz diorite; open circles: main quartz diorite with chlorite overprint; star: K-feldspar granite; triangle: biotite granodiorite.

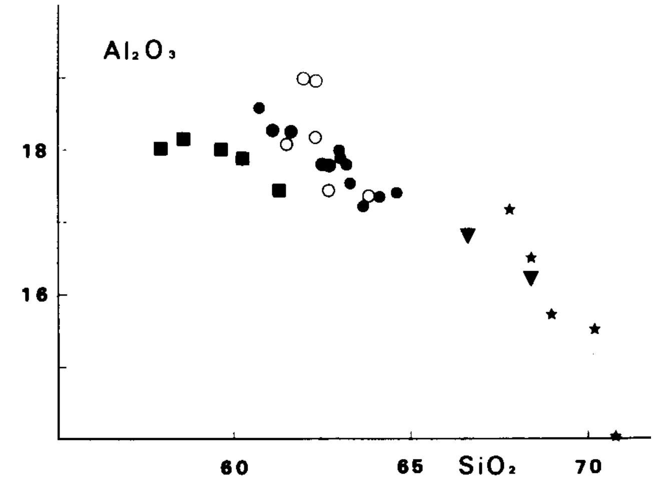

Figure 3. Al_2O_3 v. SiO_2 wt.%. Symbols as in Figure 2.

classified as peraluminous with the notable exception that most of the Fort Tourgis type are metaluminous. However there is a sharp rise in A/CNK values for a small range (60-65%) of SiO_2 content for the quartz diorites and it is probable that some of this spread may be correlated with the development of chlorite in the rock. It should be emphasised that this type of plot may be misleading when applied to rocks that have undergone later alteration.

On binary plots of the major element oxides against SiO2 the rock suite superficially shows linear trends from the more basic Fort Tourgis type to the K-feldspar granite. However a more detailed inspection suggests a more complex distribution. The K-feldspar granites are separated from the others by a compositional gap in SiO2 content. However the gap could be considered to be partly bridged by the biotite granodiorite samples. The Fort Tourgis specimens often lie on a distinctly different trend to that of the main quartz diorites and the quartz diorites sometimes show rather a scatter of points which may in part reflect late alteration events. In the plot of A1203 against SiO2 (Fig. 3) some of these features are particularly evident.

Trace elements

Trace element binary plots against SiO_2 often discriminate clearly between the three lithological groups. Generally those samples from Fort Tourgis occupy a distinct field. Those from the quartz diorite either cluster closely together without much variation or for certain elements (La, Sc, Sr) show a rather disproportionate range of concentrations. The K-feldspar granite samples usually have slightly lower values than those of the quartz diorites.

The Fort Tourgis lithology contains significantly higher Cr, Ni and V than the other samples. Figure 4 shows the distribution for vanadium but plots for chromium and nickel have an almost identical form. The Fort Tourgis lithology is dramatically discriminated on the cerium plot

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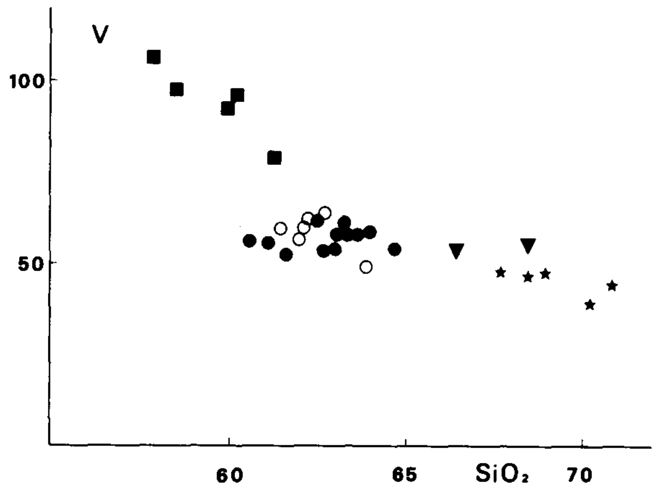

Figure 4. V ppm v. SiO_2 wt.%. Symbols as in Figure 2.

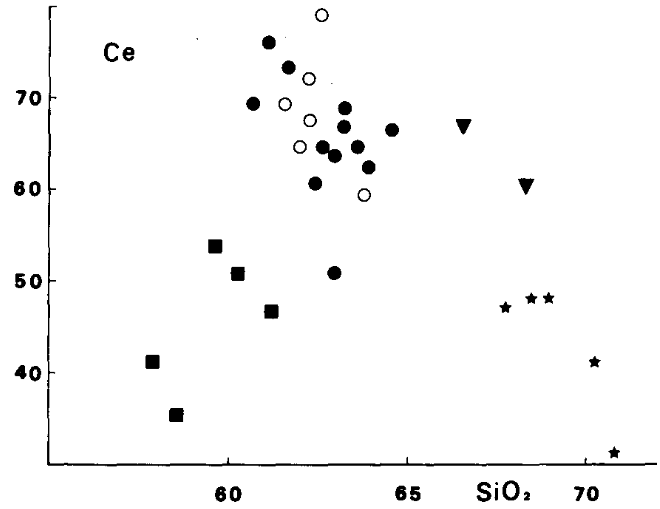

Figure 5. Ce ppm v. $Si0_2$ wt.%. Symbols as in Figure 2.

(Fig. 5) forming a field of low values and lanthanum shows a similar distribution. Y on the other hand, is highest in the Fort Tourgis samples and they show a distinctive trend on a plot of yttrium against zirconium (Fig. 6). Zirconium only shows a restricted range and is lowest in the K-feldspar granite. Strontium is highest in the Fort Tourgis samples and shows a rather disproportionately wide range in the quartz diorites (Fig. 7).

Rare earth element, chondrite normalised, profiles (Fig. 8) display relative enrichment in the light rare earths and have no europium anomalies. The K-feldspar granites are depleted in rare earths compared with the other samples analysed. The Fort Tourgis samples are depleted in light and enriched in heavy rare earths relative to those from the quartz diorites. Thus a ratio such as chondrite normalised La to Yb or equally a simple La to Y ratio forms an effective means of discriminating the Fort Tourgis samples. They have La/Y values less than two whilst all the other samples have values greater than two. Potassium and scandium, contents seem directly related to chlorite development. A plot of potassium against scandium (Fig. 9) places nearly all the chlorite bearing samples in the low field defined by 2.5%K and lOppm Sc. The Fort Tourgis samples also plot in this field. In addition this plot shows that the K-feldspar granite samples do not contain significantly more potassium than the quartz diorites with large biotite flakes. The higher Sc contents could be present in this biotite.

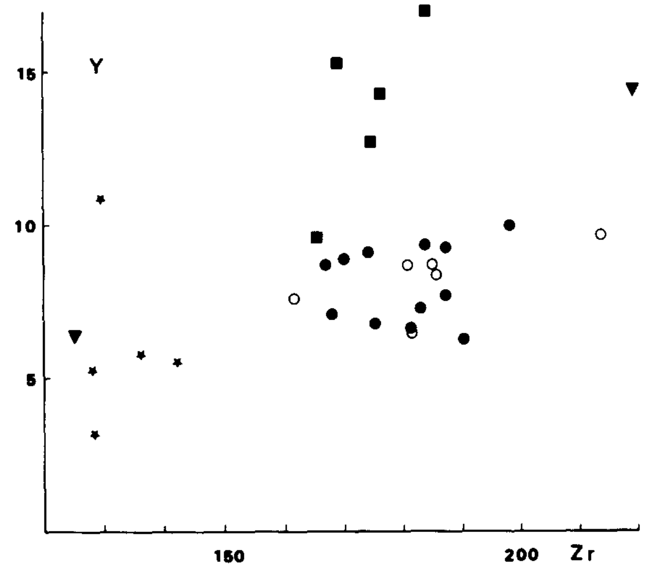

Figure 6. Y ppm v. Zr ppm. Symbols as Figure 2.

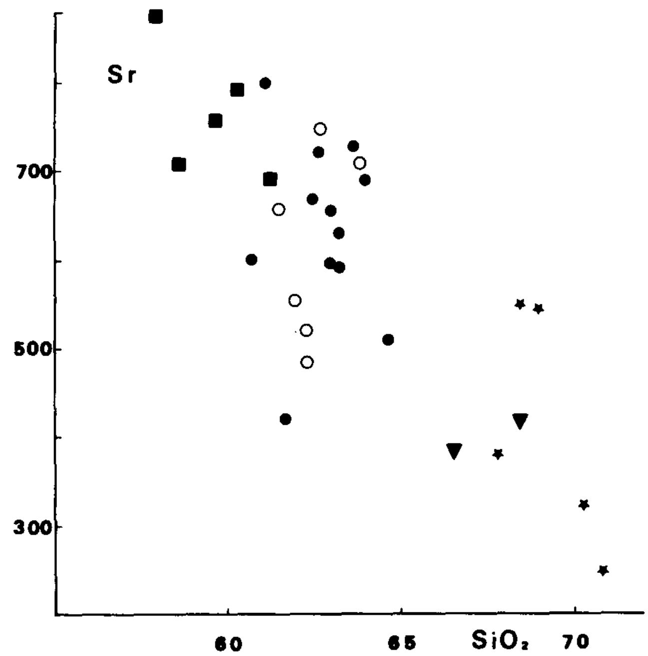

Figure 7. Sr ppm v. SiO_2 wt.%. Symbols as in Figure 2.

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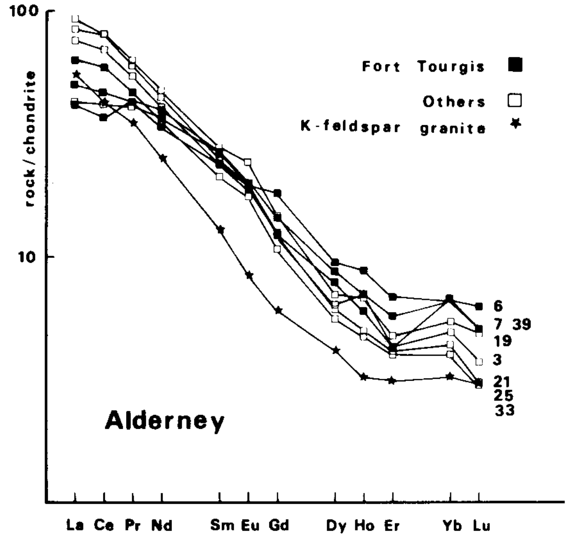

Figure 8. Rare earth element chondrite normalised profiles.

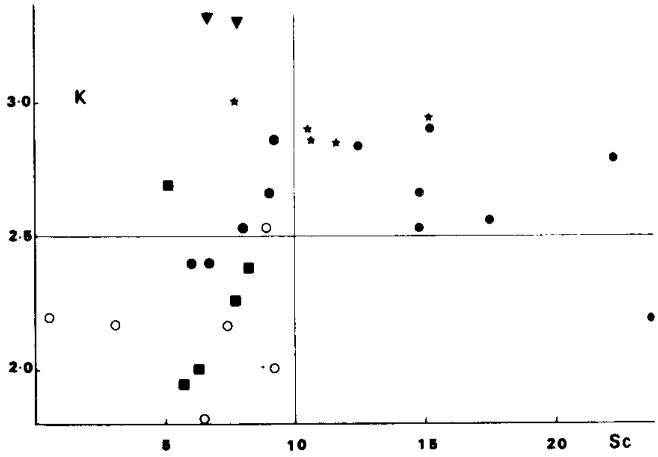

Figure 9. K wt.% v. Sc ppm.

Discussion

The foliated granitic rocks of Alderney represent a fairly complex geochemical system and there are a number of potential influences on their final chemical composition. These include not only magmatic processes and possible contamination of the magma but also later alteration. It is perhaps not possible to isolate with any certainty the effects of the various processes but the three distinctive units established within a single pluton are a useful basis for discussion. The quartz diorites are the most abundant rock type and they form a convenient reference point with respect to the other two units. However it should be recalled that there are no sharp boundaries between any of the units. The quartz diorites show a rather restricted range of chemical composition for many elements and even those samples with chlorite replacing hornblende and biotite generally plot within the same field as the other quartz diorites and only show an increased scatter for some elements. Whilst the formation of chlorite does not seem to have resulted in any major modification of the chemistry it is not possible to assess the effect of the growth of biotite on the original chemistry of the quartz diorites because of the difficulty in determining how much of the biotite is secondary. The textural evidence does suggest that the growth of biotite from hornblende would require at least the introduction of potassium into the system. The K-feldspar granites have an intimate field relationship with the quartz diorites and the most likely possibilities for their origin would seem to be either the result of local magmatic fractionation or of metasomatism. The evidence is equivocal. Certainly the K-feldspar megacrysts appear to be overgrowing the groundmass and locally they seem to be related to the emplacement of the aplite veins but if the metasomatism was purely the result of the growth of K-feldspar in pre-existing quartz diorites it is difficult to explain why the rock as a whole is much lower in mafic minerals which would be more likely to be the result of fractionation. On the geochemical plots the K-feldspar granites could be said to lie on a continuation of the same general trend as the quartz diorites again supporting a fractionation origin. However against this there is a marked gap is silica content between the rock units and in addition none of the trace elements show any relative increase with increasing silica content and the transition elements (e.g. V see Fig. 4) show remarkably little decrease in the K-feldspar granites compared with the quartz diorites.

It is tempting to ascribe the very distinctive chemistry of the Fort Tourgis type to simple contamination of a quartz diorite magma by a more basic component. The trend of the Fort Tourgis samples on many of the plots suggest mixing curves between quartz diorite and a more basic component and the presence of shadowy enclaves and xenocrystic hornblende also support this interpretation. If this is what created the distinctive chemistry of the Fort Tourgis zone it is not clear whether this was a closed system process possibly between cumulates and magma or an open system process between magma and contaminants of an undefined composition.

Although the trace element geochemistry of the quartz diorite has a restricted range of chemical composition certain trace elements have proved to be powerful discriminants. The rare earth elements and the La/Y ratio have been demonstrated to be extremely useful in recognising rocks of the Fort Tourgis type as have the transition elements Cr, V and Ni, whilst Sc and K levels were valuable in the recognition of rocks containing replacive chlorite.

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Read at the Annual Conference of the Ussher Society, January 1986

Ore genesis at Wheal Pendarves and South Crofty Mine, Cornwall - a preliminary fluid inclusion study

R.C. SCRIVENER
T.J. SHEPHERD
N. GARRIOCH

Scrivener, R.C. Shepherd, T.J. and Garrioch, *N. 1986.* Ore genesis at Wheal Pendarves and South Crofty Mine, Cronwall - a preliminary fluid inclusion study. *Proceedings of the Ussher Society 6, 412-416.*

Microthermometric data from fluid inclusions in quartz from Wheal Pendarves and South Crofty Mine are interpreted in relation to the mineral paragenesis in the veins and wallrock. The presence of an early high-salinity fluid of magmatic origin is recognized and there is some evidence of liquid immiscibility events in the subsequent development of the hydrothermal mineralisation. The influence of mixing with calcium chloride brines at a relatively early stage is recognized in the South Crofty veins.

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Introduction

In south-west England the metalliferous veins may be considered in two broad groups: those bearing tin and tungsten within and close to the granites and those in the mainly sedimentary host rocks in which base metal sulphides predominate. Veins in and around the granites are commonly enclosed within tourmaline-feldspar and/or greisen envelopes. The sulphide veins with gangue assemblages of chlorite and spar minerals commonly have extensive argillic wallrock alteration. Accounts of the structure and zoning of mineral deposits have been furnished by Dines (1956) and Hosking (1964). In this paper samples are reported on from South Crofty Mine and from Wheal Pendarves.

South Crofty Mine, which includes parts of the former Dolcoath, Tincroft and East Pool and Agar setts, is situated in the north-eastern part of Cambome between Tuckingmill and Illogan Highway. The veins trend ENE-WSW and mostly dip steeply towards the north: NNW-SSE crosscourses are present and cut the metalliferous structures. Contact - metamorphosed Upper Devonian sedimentary rocks (Mylor Slate Formation) and greenstones comprise the host rock at surface. The north-dipping contact of the Cam Brea granite ridge trends ENE-WSW, and granite is the host rock at the deeper levels of working. Granite-porphyry ("Elvan") dykes cut both granite and the sedimentary rock host. Copper and arsenic were formerly worked from the upper levels of the mine, and tin and, formerly, some tungsten from the lower. Current production (1985) runs at some 1,800 tonnes tin metal in concentrates per year.

Wheal Pendarves has exploited extensions of lodes formerly worked in the Wheal Harriet and Wheal Tryphena properties. It is situated to the south-west of Camborne, with the main shaft some 1.2 km due east of Barripper Church. Most of the development is on two structures of similar trend to those at South Crofty, Harriet Lode to the north and Tryphena Lode to the south. Both structures lie at depth in an extension of the Cam Brea Granite while the upper levels of Tryphena lode are in slates. Lodes in the vicinity of Wheal Pendarves were formerly worked for copper; at present the production is of tin.

Methods

The study is based on the examination of thin and polished sections of veinstone and wallrock from the mines and on the microthermometric analysis of 20 doubly polished wafers of quartz.

Microthermometric analysis was carried out using a Linkam TH600 heating-freezing stage and control unit. The results of the heating-freezing experiments are quoted as TH, the temperature of homogenisation - usually marked by disappearance of a vapour bubble -and salinity - measured from the last ice melting temperature or the temperature of halite solution. No pressure correction has been applied to TH data, which must therefore be regarded as minimum trapping temperatures. Salinity is expressed as weight % sodium chloride equivalent except where low eutectic temperatures indicate the presence of calcium chloride; in this case the hydrohalite melting temperature may be used to estimate the relative proportion of calcium chloride to sodium chloride.

Vein mineralogy and paragenesis

At South Crofty Mine, the structures examined fall into three categories:

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1. *'Pegmatitic' lodes.* These are early quartz-feldspar bodies, commonly flat-lying, with wolframite and lollingite or arsenopyrite. They are commonly cut by a later generation of quartz-tourmaline-cassiterite veins. Material from the Complex and 3ABC lodes was examined.

2. *'Normal' lodes.* These are steeply dipping quartz-tourmaline-chlorite-cassiterite-sulphide veins, which commonly show brecciation and evidence of repeated opening. Material from Pryces, North and Tincroft lodes was collected for microthermometric analysis. Thin sections from North Lode show several stages of tourmaline-quartz and tourmaline-cassiterite-quartz growth with later generations, commonly multiple, of quartz-fluorite-chlorite and quartz-hematite. The principal cassiterite mineralisation occurred at an early stage and is represented by rounded aggregates of cassiterite with a little tourmaline and quartz in a finely crystalline mass of tourmaline with quartz, minor chlorite and scattered anhedral crystals of cassiterite.

3. *'Crosscourses'.* These narrow, vertical or subvertical quartz, quartz-fluorite or quartz-hematite veins cut across the tin-bearing veins throughout the workings. They are generally banded, with fluorite present in the earlier stages of mineralisation. Wallrock alteration, apart from local argillisation is not well-marked.

At Wheal Pendarves the two main mineralised structures are contrasted in structure and paragenesis. Harriet Lode is a sheeted complex of quartz-tourmaline-cassiterite veins with some later sulphide-fluorite, in an envelope of altered granite. The wallrock alteration involves tourmalinisation and the growth of secondary potassium feldspar in the early stages, and chloritisation and argillic alteration in the late stages. Only one stage of cassiterite deposition was noted in the Harriet veins; this was preceded and succeeded by the growth of tourmaline.

The Harriet mineralisation is considered to be earlier than that in the Tryphena Lodes, in which cassiterite was deposited at an early stage in the mineralisation, and may be predated by quartz in association with hematite. Chlorite-quartz-sulphide (including arsenopyrite) postdate the cassiterite-quartz stage and may be associated with fluorite and hematite. Late quartz is ubiquitous and may be associated with goethite and other secondary minerals. The Tryphena veins are notably deficient in tourmaline, and the wallrock alteration is argillic.

Fluid inclusion studies

Microthermometric examination of material from both mines proved to be difficult; primary and pseudosecondary inclusions amenable to analysis are scarce, and most of the wafers showed many small-scale healed fractures with well developed curtains of very small secondary inclusions. In some cases the secondary inclusion trains have completely obliterated earlier inclusions and the best wafers yielded only a small number of possible analyses. A further difficulty was experienced in the examination of wafers cut from material blasted with modern explosives. Extensive microfracturing took place in them during the high temperature part of the heating - freezing cycle, a feature which greatly reduced the number of homogenisation events observed.

Four types of inclusions were observed:

a) Liquid only. Very common as secondary inclusions.

b) Liquid-vapour. These mostly small, equant to irregular inclusions are the most common of those measured. Necking-down of two phase inclusions is very common especially in Complex and 3ABC lodes of South Crofty.

c) Liquid-vapour-halite. Three phase inclusions are present in all the structures examined except the Pendarves Tryphena veins and the crosscourses, but they are scarce and generally very small. In every case where measurement of TH was possible, homogenisation was by disappearance of the vapour bubble. Very rare, small, irregular inclusions with more than one daughter salt were noted in quartz from Pryces Lode and in wallrock from North Lode. A group of inclusions in quartz from 3ABC lode and from Pendarves Harriet wallrock bore 3 or 4 solid phases plus vapour and liquid. These decrepitated at temperatures in excess of 375° before solution of the halite phase: the salinity of these inclusions must be in excess of 40 wt% NaC I equivalent.

d) Vapour-rich. Vapour filled inclusions are rare in the material examined, though a small population associated with high TH liquid-vapour inclusions was noted in quartz-cassiterite from North Lode, South Crofty.

A plot of TH against salinity for all South Crofty data, excluding crosscourses, is shown in Figure 1. There is a broad scatter with TH values ranging from 123° to 344°C and salinity from < I to 35 wt% NaC I equivalent. The graph shows a group of low salinity (< 5 wt% NaC 1) fluids ranging in TH up to 230°C, probably representing a late, low salinity event. Quartz intergrown with cassiterite in North and Pryces lodes and quartz intergrown with K-feldspar and wolframite in Complex and 3ABC lodes both have a population of inclusions in the TH range 250-350°C and salinities from 4-32 wt% NaCl equivalent. A small number of low salinity, high - TH inclusions from Complex Lode showed evidence of CO_2 content by clathrate formation at sub-ambient temperatures. Note was taken of any inclusions showing eutectic melting in the range -60°C to -50°C, which is an indication of the presence of $CaCl_2$ A group with this characteristic is present in both North and Pryces lodes. A relatively small number of high - salinity multiphase inclusions cannot be shown on Figure 1. due to premature decrepitation and/or microfracturing of the host quartz at elevated temperatures. Quartz associated with deep purple fluorite was collected from a

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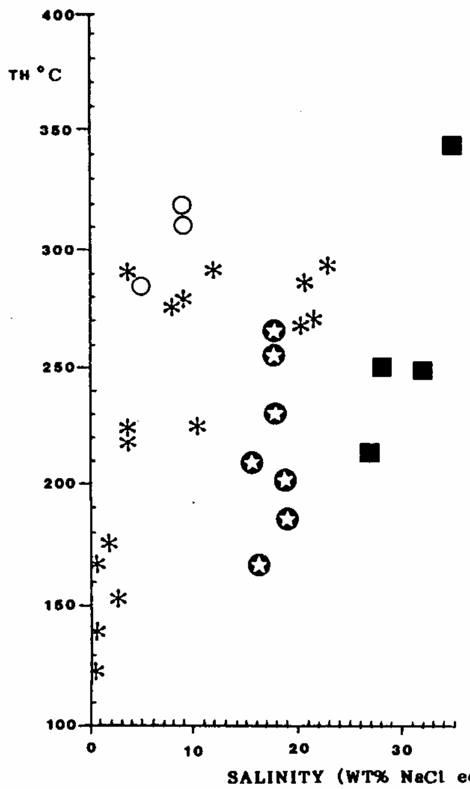

FIG.1. TH-SALINITY PLOT FOR ALL SOUTH CROFTY DATA EXCLUDING CROSSCOURSES.

INCLUSION TYPES: LIQUID-VAPOUR ✱
LIQUID-VAPOUR-HALITE ■
CaCl2 - RICH ✪
CO2 - RICH ○

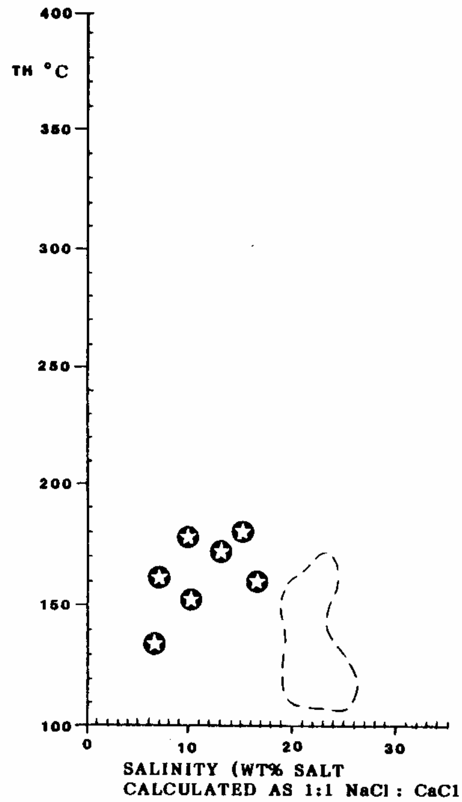

FIG.2. SALINITY PLOT FOR CROSSCOURSE DATA, SOUTH CROFTY MINE. ALL INCLUSIONS ARE LIQUID-VAPOUR. DASHED LINE SHOWS FIELD ENCLOSING 35 DATA POINTS FOR CROSSCOURSE INCLUSIONS IN THE TAMAR VALEY OREFIELD (E. CORNWALL).

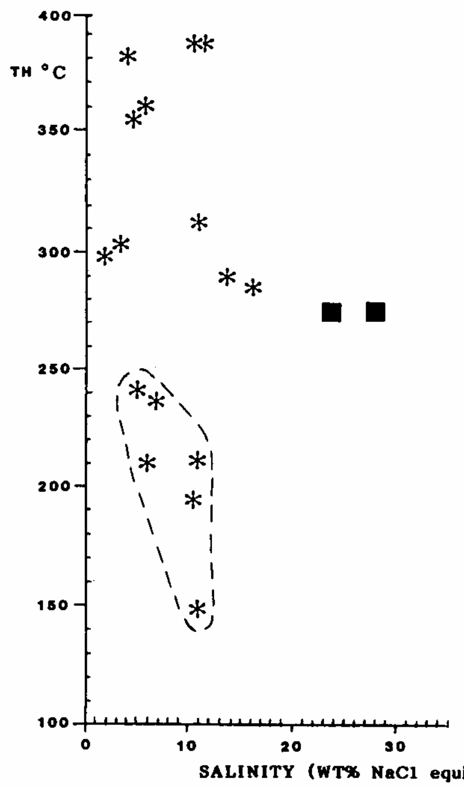

FIG.3. TH-SALINITY PLOT FOR ALL WHEAL PENDARVES DATA. KEY TO INCLUSION TYPES AS FOR FIG.1. TRYPHENA DATA IS ENCIRCLED BY DASHED LINE.

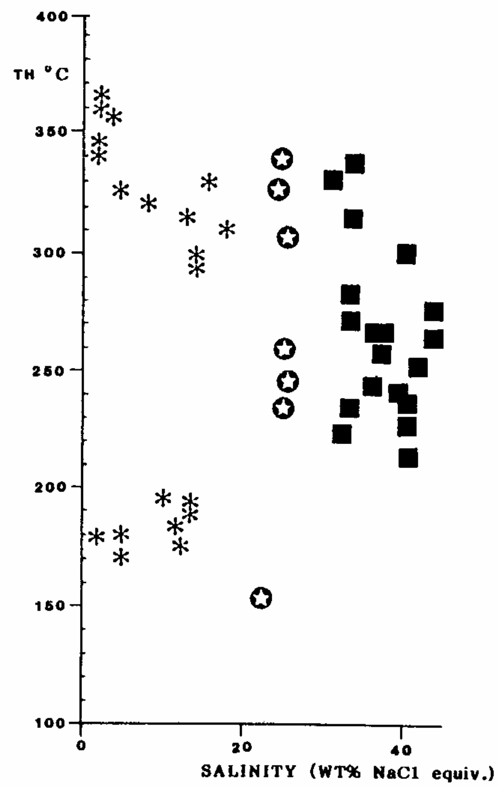

FIG.4. TH-SALINITY PLOT FOR INCLUSIONS IN QUARTZ FROM WALL LODE BIRCH TOR & VITIFER MINE, DARTMOOR. KEY TO INCLUSIONS AS FOR FIG.1.

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crosscourse that intersects Pryces Lode, and the fluid inclusion data are presented in Figure 2. TH ranges from 132°C to 180°C and salinities from 7 to 18 wt% NaC I equivalent. In all cases a eutectic between -65°C and -50°C was observed and the presence of $CaCl_2$ in the brines was inferred. Hydrohalite melting temperatures for the more saline inclusions indicate that the $CaC1_2$ to NaCl ratio is about 1:1.

A TH-salinity plot (Fig. 3) for the Harriet two phase (liquid-vapour) and three phase (liquid-vapour-halite) data give a range of salinities from 1.8 to 28.0 wt% NaCl equivalent, and a range of TH from 275°C to 385°C. Homogenisation in all cases was by disappearance of the vapour bubble. Multiphase inclusions are relatively common, especially in the tourmalinised wallrock, but microfracturing at elevated temperatures precluded measurement of TH. There is no doubt that early high-salinity fluids are important in this system.

In contrast to the relatively high TH values for the Harriet system with its spread of salinities and inclusion types, quartz-cassiterite-chlorite intergrowths from the Tryphena veins show (Fig. 3) a lower TH range of 150°C to 240°C and salinities restricted between 6.0 and 11.1 wt% NaC I equivalent. The small number of two phase liquid-vapour inclusions are masked by finely divided chlorite, rendering some wafers nearly opaque, and by extensive networks of secondary inclusions along healed microfractures. Only six data points were realized from 3 wafers.

Discussion and conclusions

The data presented in this report, when taken together with earlier work of Bull (1982), Scrivener (1982) and Shepherd and others (1985), support the recognition of two types of ore genesis. These may be summarised as follows:

1. Tin-tungsten mineralisation associated with extensive early tourmalization and pegmatite development.
2. Tin-base metal sulphide mineralisation associated with chlorite-fluorite gangue assemblages and extensive argillic alteration.

The Harriet mineralisation at Wheal Pendarves falls in category 1 above, while 2 is typified by the Tryphena system at Wheal Pendarves. The mineralisation at South Crofty comprises a combination of types 1 and 2. The structures in which type 1 predominates being earlier than those dominated by 2.

Category 1 is the result of deposition from initially highly saline fluids possibly of magmatic origin. Although the TH range of most of the inclusions with salinity 12 wt % NaCl equivalent is from 200° to 300°C and rather lower than might be expected for a magmatic provenance, the lack of evidence of boiling and the paucity of vapour rich inclusions do not support alternative mechanisms for their origin. There is some evidence *(cf.* Shepherd and others 1984) of liquid immiscibility giving rise to a high-temperature low-salinity CO_2-rich phase that is responsible for the wolframite mineralisation. Characteristically, the fluids responsible for category 1 mineralisation show a great range in salinity and TH. There is little evidence of a correlation between Th and salinity and it appears that high and low salinity events occurred within the same temperature range. A comparison of data from Harriet mineralisation at Pendarves and from South Croft', with results from a typical tin-tourmaline-haematite-quartz vein from the Birch Tor district (Central Dartmoor) (Fig. 4) shows similar spreads of TH and salinity. Both the Birch Tor and South Crofty examples show fields of low-TH and low-salinity inclusions that represent late gangue deposition events. At both Birch Tor and in the South Crofty normal lodes (Fig. 1) there is evidence that cross-courses (calcium chloride-rich) brines entered the system at an early stage and may have influenced tin deposition. The Harriet system does not show this characteristic and appears also to lack a low TH, low salinity stage.

Type 2 mineralisation is from low to moderate salinity fluids of wide-ranging TH. Vapour- rich inclusions are absent and there is no evidence of boiling in the Pendarves Tryphena veins. The absence of low eutectic melting in these inclusions suggests that calcium chloride-rich brines are not important, though there is insufficient data available to make this conclusive. The work of Bull (1982) suggests that in cassiterite-base metal sulphide mineralisation in East Cornwall and West Devon, salinity decreases with temperature through paragenetic time. Thus, early cassiterite deposition from low to moderate salinity fluids at high TH is followed by sulphide deposition at low salinity and low TH. Further work, particularly on sulphide deposition, would be necessary to establish this model in the present district.

Acknowledgements Thanks are due to the management and staff of the two mines and to colleagues who commented on the manuscript and provided helpful discussion.

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Read at the Annual Conference of the Ussher Society, January 1986

'Lake Bude' (early Westphalian, SW England): storm-dominated siliciclastic shelf sedimentation in an equatorial lake

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It is widely accepted that the Bude formation is lacustrine (Goldring and Seilacher, 1971; Higgs, 1984), although the precise environment is controversial (Higgs, 1983, 1984). The lake, which lay within a few degrees of the palaeomagnetically-determined equator (Scotese *et al.,* 1979), is here named Lake Bude. The lake was initiated by isolation of the pre-existing Crackington Formation E-W seaway in late Namurian time (Freshney and Taylor, 1972; Freshney *et al,* 1979) and was terminated when the N-advancing Variscan mountain front swept across the area in late Westphalian and/or Stephanian time (Shackleton *et al.,* 1982). Apart from three brief marine episodes (Freshney *et al.,* 1979), the water in Lake Bude alternated between fresh and brackish (see below); this suggests that the lake was separated from the open sea by a sill which was occasionally overtopped by sea water, perhaps due to glacioeustatic sea-level fluctuations. This scenario is closely comparable to the late Quaternary history of the Black Sea (Scholten, 1974).

The Bude Formation consists of about 1300m of mudstones and intercalated very-fine sandstones (= event beds) (Higgs, 1984). Many of the event beds are capped by wave-influenced ripples *(op. cit.);* this suggests that the 'events' were characterized by combined flows consisting of a unidirectional (sediment-supplying) current and a wave-induced oscillatory current. Most authors agree that the Bude Formation was deposited entirely offshore *(op. cit)* corresponding shorelines deposits are unknown. (The early Westphalian paralic Bideford Formation of N Devon (Edmonds *et al.,* 1979) is thought to be older than all but the lowermost few tens of metres of the Bude Formation (Higgs, 1986). It is argued below that the Bude Formation accumulated on a lake shelf, which passed southward into an inferred flysch trough. Sole marks indicate that (sediment-supplying) event currents flowed from all quadrants except the south.

The succession shows a pervasive dm-m scale cyclicity consisting of an alternation of sand-rich and sand-poor intervals. A symmetrical 'ideal' cycle can be defined, comprising three facies associations (FA), arranged in the following order: (FA1) dark grey laminated mudstone (cm-dm) with a few thin (mm-cm) event beds + (FA2) light grey silty mudstone (dm) with one or two thicker (up to 30 cm) event beds 3 (FA3) a thick (dm-m) amalgamated sandstone unit, located in the middle of the cycle, consisting of relatively *thin* (mostly 10-30 cm) event beds (FA2) as above (FA 1) as above.

FA1 and FA2 contain body- and trace fossils which yield important palaeosalinity information. Body fossils occur in FA I mudstones: apart from the three marine horizons (cm) with goniatites and pelagic bivalves, they are limited to rare fish and crustacea (Freshney *et al.,* 1979). Burrows in FA1 include *Planolites* (King, 1965), and 'post-event' U-burrows *(Arenicolites* and *Diplocraterion)* which are confined to the tops of event beds. The total faunal assemblage in FA is interpreted as indicating that deposition took place in brackish, dysaerobic (Rhoads and Morse, 1971) water. FA2 contains no body fossils or burrows, but has yielded king-crab trackways (King, 1965), as well as sinusoidal traces *(op. cit.)* which are here interpreted as fish trails (Higgs, 1986); based on this assemblage and the absence of burrows, FA2 is thought to have been deposited in fresh, aerobic water. Sulphur and organic-carbon analysis of five FA2 silty mudstones yielded C/S ratios ranging from 20 to 40; these values support the fresh-water interpretation (Berner and Raiswell, 1984). FA3 contains neither body- nor trace fossils; a fresh-water environment is inferred, since FA3 is sandwiched by FA2 in the ideal cycle.

It appears, then, that the ideal cycle reflects a change from brackish- to fresh- and back to brackish-water conditions. At the same time there is a decrease then an increase in *depth,* as evidenced by the fact that the muds of FA2 were siltier and better oxygenated (see above) than those of FA 1; the cyclicity is therefore interpreted as a transgressive-regressive cyclicity (cf. Busch and Rollins, 1984). The salinity fluctuations suggest that the cyclicity was externally controlled (i.e. allocyclic).

The event beds are interpreted as `underflowites', deposited by river-fed underflows during catastrophic storm-floods: waning of the underflows allowed simultaneous storm-wave action to rework the tops of many beds (Higgs, 1986). FA1, FA2 and FA3 are all thought to have been deposited above storm wave base, since wave-influenced event beds are present in each.

In view of the evidence for a relatively shallow

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environment (i.e. above storm wave base), it is peculiar that the cycles lack a nearshore/emergent component. One possible explanation is that deposition took place on a storm-dominated shelf, and that aggrading fair-weather sediment was winnowed down to a storm-wavecontrolled 'equilibrium surface' each time a storm occurred, the surplus sediment being swept over the shelf edge into deeper water (cf. Seilacher, 1982, p.171). This model implies that there must have been a neighbouring trough in which the surplus sediment was accommodated. The inferred trough, of which there is no exposed/preserved sedimentary record (due to over-thrusting?), is thought to have lain in the S in the form of a flysch trough occupying the S part of Lake Bude; this would explain the lack of N-flowing palaeocurrents in the Bude Formation, and would imply that the S shoreline of the lake was coincident with the (advancing) Variscan front.

In the context of the proposed shelf model, the so-called "slumped beds" (Freshney *et al,* 1979) are reinterpreted as seismites (cf. Seilacher, 1969), formed *in situ.* This interpretation accords with the following observations: (i) soft-sediment folds within the "slumped beds" show no strongly preferred orientation (Whalley and Lloyd, 1986); and (ii) "slumped beds" pass gradationally into uncontorted sediment when traced laterally (Melvin, 1986).

FA2 silty mudstones commonly show, in thin section, a disturbed horizontal lamination, suggestive of disruption by upward-migrating gas bubbles (presumably methane) (cf. Fig. 363 of Reineck and Singh, 1980). This evidence for methane generation at very shallow (cm-dm?) burial depths is consistent with the fresh-water interpretation advocated above, since methane production at such shallow depths is possible only in fresh-water environments (Curtis, 1977). It is proposed, therefore, that during fresh-water episodes the surficial muds were full of gas bubbles. It follows that the muds would have been exceptionally prone to resuspension during storms; this may explain why the thick (up to 10 m) central amalgamated unit (FA3) of the Bude Formation cycles invariably consists of relatively *thin* (max. 40 cm) event beds; in other words, each event may have caused the removal of mud which, under normal (i.e. marine) circumstances, would have stayed in place. One might speculate that if the lake had instead been a (marine) sea, the ideal cycle would have consisted of a gradually thickening-upward/thinning-upward sequence of *non-amalgamated* event beds.

Acknowledgements. This note documents the main conclusions of my doctoral research on the sedimentology of the Bude Formation. I thank Dr. H.G. Reading for his stimulating criticism throughout my studies at Oxford, and for suggesting improvements to the manuscript. The project was financed by a British Petroleum studentship and a government O.R.S. award.

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Read at the Annual Conference of the Ussher Society, January 1986

Observations of the Padstow Confrontation, north Cornwall (Abstract)

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The Padstow Confrontation has been held to be a structural line of considerable importance in the tectonic evolution of the Hercynides of south-west England. On the north Cornwall coast, a large scale south-facing over-fold was modelled (eg. Hobson and Sanderson, 1975) which predicted south-facing structures as far south as Polzeath (SW 937787). North-facing folds are documented from Padstow southwards.

Roberts and Sanderson (1971) suggested an early phase of deformation, involving movement towards the north, which affected only those rocks now represented south of Polzeath. The structures to the north were attributed to later, southward directed movements which caused local refolding of the early structures at the facing confrontation. It followed that slaty cleavage to the north was related to a younger phase in the deformation history from that in the south.

In their detailed arguments, the Gravel Caverns Section, Polzeath occupies a critical position. The only DI structure recorded was an S I fracture cleavage which was reported to be refolded about north-verging but south-facing F2 folds bearing a well developed north-dipping slaty (S2) cleavage. A position on the inverted limb of a regional D2 overfold was deduced. Our investigations in this section show that SI is a well developed slaty cleavage, axial planar to M-, S- and Z-shaped minor folds which accompany larger scale tight recumbent F1 folds. The implication is that this fold style is regional. The F1 structures which face north, are associated with a 20 m thick flat-lying thrust zone on which the Gravel Caverns Conglomerate overthrusts the Polzeath Slate *(cf.* Gauss and House, 1972).

At Gravel Caverns these D 1 structures are refolded by north-verging F2 folds with an axial planar south-dipping spaced cleavage (S2). Such folds generated downward- and even south-facing on refolded SI cleavage represented in their steep and overturned limbs. An identical structural style can be identified, not only in the Pentire Slate section on the foreshore north of Gravel Caverns, but throughout coastal sections extending for many kilometres to the north. Everywhere, the direction of tectonic transport in D I and D2 is northwards, S I is always a slaty cleavage associated with DI, and S2, a locally developed, south-dipping, D2 spaced cleavage. The case for the confrontation of fold facing fails.

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British Triassic palaeontology: supplement 10

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A wide range of organisms is known from Triassic deposits in the British Isles but relevant literature is widely dispersed and in some cases is obscure; recognition of these factors prompted the preparation of an earlier review by the writer *(Proceedings of the Ussher Society, 3,* 341-353; 1976). Relevant publications have continued to appear, however, and that account has been updated with bibliographic supplements published annually in the *Proceedings of the Ussher Society.*

Since 1976 relevant contributions have appeared in 48 serials ranging from *Acta Geologica Polonica* to the *Zoological Journal of the Linnean Society,* in addition to non-serial works; about one third of the titles are published outside Britain.

In the nine supplements published previously, and that appended here, 178 separate contributions are listed. Growth in the number of items cited, from four in the first supplement to 21 or more in recent issues, and in the number of items published each year, from four in 1976 to 17 or more in most subsequent years, indicates that interest and activity in the subject area has expanded considerably during the period under review. Since the completion of the writer's previous supplement *(Proceedings of the Ussher Society, 6,* 273; 1985) to his paper on British Triassic palaeontology, the following works relating to aspects of that subject have been published or have come to his attention; some earlier works, including items omitted from the original review and earlier supplements, are included:

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

Read at the Annual Conference of the Ussher Society, January 1986

Lithological and geochemical characteristics of some Gramscatho turbidites, south Cornwall (Abstract)

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The compositions of sandstone turbidites from the Gramscatho Group of south Cornwall have been determined to establish provenance and aid identification of the plate environment of emplacement. Petrographic analysis shows that the framework grains are quartz, feldspar and lithic fragments in approximately equal proportions. Acid volcanics predominate in the lithic component but the significant proportion of metamorphic grains increases up the sequence. The QFL and complimentary triangular plots of grain sub-populations all indicate a dissected continental magmatic arc provenance for the sandstones and suggest emplacement in a fore-arc setting.

The sandstones are chemically classified as quartz-intermediate greywackes with SiO_2 contents of about 70 wt.% and a major element composition not dissimilar to other Phanerozoic greywackes. K, Rb and Cs abundances are variable and reflect variable detrital alkali feldspar rather than superimposed metasomatism. Ratios involving La, Th, U, Ni, Cr and REE are essentially constant, establishing the chemical uniformity of the greywackes and confirming derivation from an upper continental source of predominantly acidic composition. Chemical discrimination of the depositional environment, in agreement with the clast mode data, indicates accumulation in a basin adjacent to a continental magmatic arc.

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Manuscripts and enquiries on editorial matters should be directed to the Editor at the following address:

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