"STRATIGRAPHY, SEDIMENTOLOGY AND PALAEONTOLOGY OF THE LOWER CARBONIFEROUS OF ANGLESEY"

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by

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ABSTRACT

The Anglesey Dinantian succession records the establishment and growth of a land-attached carbonate platform. Onlap of underlying basement terrain and marginal terrigenoclastic accumulations is readily demonstrated. Carbonate deposition however was repeatedly interrupted, palaeokarstic surfaces recording the periodic lowering of sea level and the emergence of extensive areas of the limestone shelf. The sequence is constructed therefore of numerous transgressive/regressive minor cycles. These have been grouped together into five broader, lithostratigraphically based formations.

During periods of raised sea level active carbonate production was achieved in a mosaic of facies of variable but not pronounced water depth. During periods of marine regression and emergence the carbonate platform was subject to the effects of subaerial weathering. Wind-blown volcanic ash accumulated on the exposed limestone surfaces and was colonised and stabilised by vegetation. Complex pedogenic alteration effects were promoted in the underlying carbonate sediment whilst dissolution beneath such soil covers led to the distinctive hummocky topography of palaeokarstic surfaces.

The lowering of erosive base level during regressive periods also resulted in the rejuvination of siliciclastic source areas within the adjacent hinterland of older rocks. Marginal alluvial fans prograded onto the emergent shelf and rivers incised complex channel systems. Beyond the marginal fans transportation and deposition of terrigenous sediment appears to have been largely confined to such channels.

Marine transgressions saw the drowning of the channel complexes

and the shutting down of terrigenous supply, and culminated in the inundation of palaeokarstic levels and the re-establishment of carbonate facies mosaics on the shelf.

CHAPTER ONE

GENERAL INTRODUCTION

1.1 AIMS AND APPROACH

The Anglesey Lower Carboniferous (Dinantian) is a mixed sedimentary sequence of carbonate and siliciclastic rocks attaining a maximum thickness of 280 m. Hitherto work on these strata has been undertaken from a largely stratigraphic point of view trying to place the succession within broad zonal schemes and to suggest correlations with Lower Carboniferous sequences outcropping on the North Wales mainland. These previous studies (Morton, 1901; Greenly, 1919; Nichols, 1962 and Mitchell, 1964) were made without a detailed understanding of recent sedimentary environments particularly in the carbonate field. Consequently ideas on the depositional environments in which these rocks were formed are restricted to general comments concerning depth of water, energy conditions, proximity to shoreline and climate.

The present study initially steered away from the wider stratigraphic aspects and applied a more sedimentological approach. Taking advantage of the great volume of recent literature detailing recent carbonate and siliciclastic deposits, this provided an opportunity to present a more detailed environmental interpretation for the sedimentary sequences encountered.

A complimentary aim was to assess sediment/fauna relationships for the various lithofacies within the succession.

The project was, therefore, constructed very much along the lines advocated by Ramsbottom whose papers (1973 and 1977) have catalysed research within the British Dinantian during the last decade. He called for a new approach to Dinantian stratigraphy integrating sedimentological and palaeontological texhniques and suggested a new basis for correlation within the system (see Section 1.3). The current investigation, therefore, provided an opportunity to assess rigorously these fresh ideas and this in turn served to rejuvinate interest in the stratigraphic aspects of the sequence.

1.2 LOCATION OF OUTCROPS AND EXTENT OF EXPOSURE

Dinantian sequences outcrop in three main areas on Anglesey termed, after Greenly (1919, p.600), the Principal Area, the Penmon Area and the Straitside Area (Fig.1).

The Principal Area (approximately 59 km^2, excluding that part of the outcrop caught up in the Berw Fault Zone, the Esgeifiog Strip of Greenly p.647) occupies a tapering tract of land across the centre of the Island. It includes the magnificent and almost continuous cliff exposures of the N.E. coast extending from Lligwy Bay in the north to Red Wharf Bay in the south. Inland exposure from this coast becomes increasingly poor as the crop is traced south-west and although there are numerous quarries marked on the O.S. maps many of these were found to have been filled.

The Penmon Area (approximately 9 km^2) occupies the extreme eastern tip of Anglesey and includes Puffin Island. Again there are superb, though precipitous, cliff sections extending from Careg Onen in the west to the eastern point Trwyn Du. Inland exposure is generally poor although there are several large extant quarries.

The Straitside Area occupies ground (approximately 20 km^2) in the south-western corner of Anglesey adjacent to the Menai Straits, but in an extension of Greenly's original definition is also taken to include outcrops of Dinantian rocks (approximately 8 km^2) between

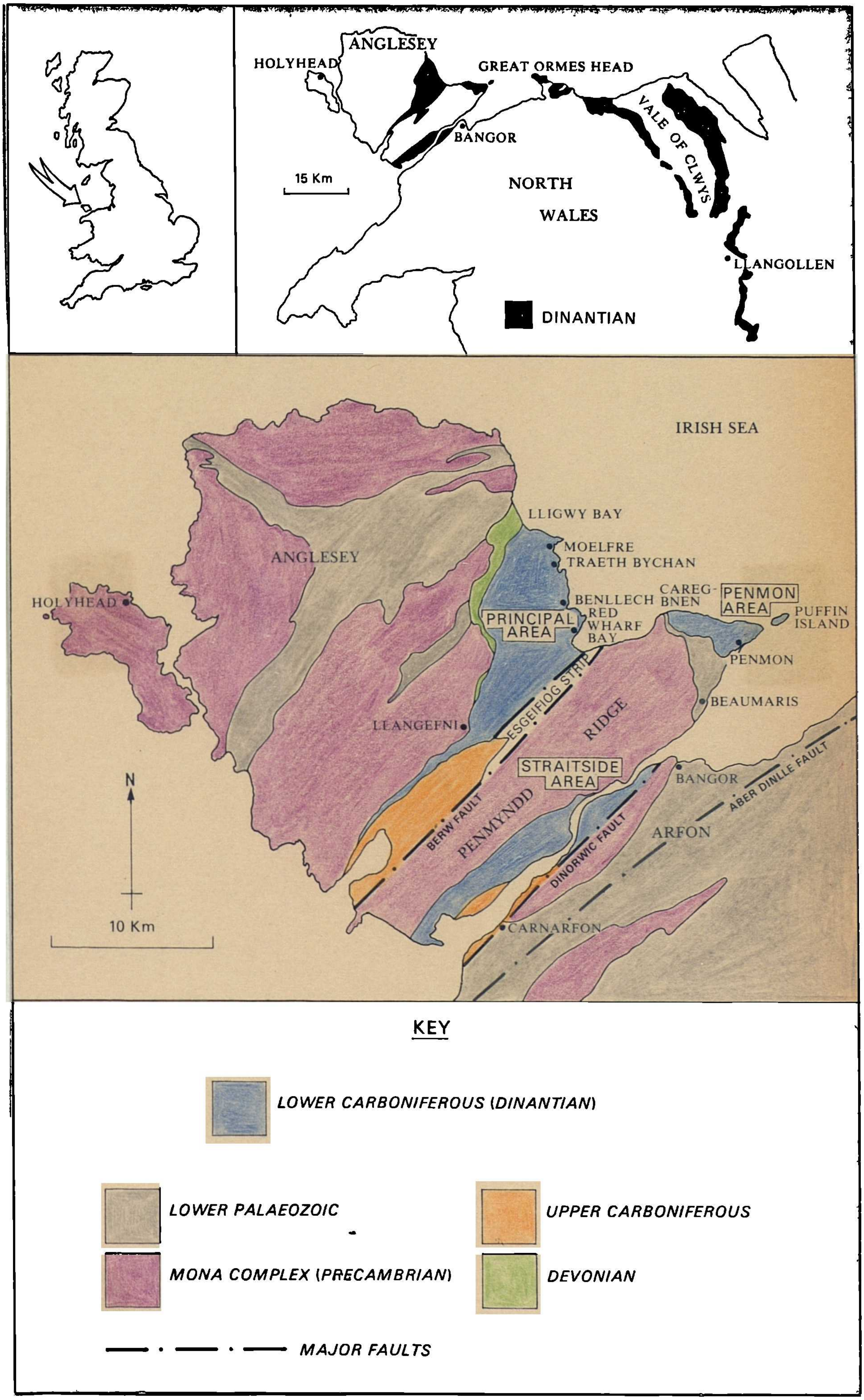


Fig.1 Geological Map of Anglesey showing main outcrops of Lower Carboniferous rocks.

Bangor and Port Dinorwic on the Arvon side of the Straits. It is the least exposed of the three areas, with exposure limited to the low, discontinuous cliff lines either side of the Straits.

1.3 DINANTIAN CORRELATION (See Fig.2 and George et al, 1976)

(a) Biostratigraphy

Until quite recently subdivisions of the British Dinantian in common usage were based on faunal zonation. In 1905 Vaughan published his classic paper on the Lower Carboniferous of the Avon Gorge near Bristol in which he established a series of faunal zones related, he thought, to progressive changes in the composition of the coral-brachiopod macrofauna. This zonation, with subsequent refinements e.g. Reynolds and Vaughan (1911); Dixey and Sibley (1918), was extended throughout the country and, although requiring many regional amendments, quickly became the basis for Dinantian correlation. Vaughan's scheme is augmented for the upper part of the Dinantian by the goniatite zones proposed by Bisat (1928) again with subsequent modification (see George et al, 1976, p.3).

During the years severe problems have been encountered in applying these macrofaunal zonal schemes, highlighted in the above phrase "regional amendments". It was found difficult to recognise the definitive Vaughanian assemblages in some parts of the country. This has generally been attributed to strong facies control of some zonal forms or marked faunal provincialism (George, 1958 and 1969), but perhaps of greater significance is the pronounced degree of non-sequence the Avonian strata-type is now known to exhibit (Ramsbottom, 1973; George, 1976). To resolve these difficulties further zonal forms and assemblages of regional significance were established and their equivalence to the Vaughanian

Fig.2 Dinantian Correlation

SUB-SYSTEM	STAGES George et al (1976)	CORAL-BRACH ZONATION Vaughan (1905)	GONIATITE ZONATION Bisat (1928)	MAJOR CYCLES Ramsbottom (1973)	MESOTHEMS Ramsbottom (1977)
DINANTIAN	Brigantian	D_2	P	6 group	D6b
					D6a
	Asbian	D_1	B	5 group	D5b
					D5a
	Holkerian	S_2		4	D4
	Arundian	S_1 C_2		3	D3
	Chadian	C_1		2	D2b
					D2a
	Courceyan	Z		1	D1c
		K			D1b
					D1a

scheme suggested (e.g. Garwood, 1907 and 1913; Hudson, 1926). The situation remains unsatisfactory. Nor are the goniatite zones of Bisat ideal. They cover only a limited part of the Dinantian succession and again display strong facies dependence being recognised with confidence in basinal and shelf edge deposits only.

Microfaunal zonal schemes are of increasing importance in the subdivision of the Dinantian. Zonation based on foraminiferal assemblages was first attempted by Cummings (1961) but was never published in full and failed to gain widespread acceptance. Recently Strank (1981) has extended the foraminiferal work of Conil and others on the Belgium Dinantian (Conil et al, 1977) to the British succession. Conodonts (Rhodes et al, 1969; Austin, 1973) and Miospores (Neves et al, 1972) also provide the basis for further biostratigraphic subdivision and are subject to increasing refinement.

(b) Chronostratigraphy

In 1973 Ramsbottom published his "new synthesis of British Dinantian Stratigraphy" in which he put forward a radical new approach to Dinantian correlation. He recognised (partly extending the early work of Dixon in Dixon and Vaughan, 1912) a cyclic arrangement of lithologies within Dinantian shelf sequences allowing their subdivision into "Major Cycles". The lithologies were interpreted as having bathymetric significance with each cycle composed of deeper water "transgressive" deposits, overlain by increasingly shallow water "regressive" phases and terminated by features indicative of emergence. Ramsbottom further suggested that these consecutive transgressive/regressive events were eustatically controlled and that recognition of the cycles therefore provided a means of chronostratigraphic correlation (Fig.2). An attractive extension to this theory is that it also provides an explanation for the Vaughanian zonal scheme. Each transgression is thought to carry with it a fresh and more highly evolved fauna. Where there are significant non-sequences between cycles as in the Avon Gorge the contrast between successive faunal assemblages is further emphasized and was consequently readily recognised by Vaughan. It follows that, in theory, the "Major Cycles" and the coral-brachiopod zones are coincident.

In an extension of his model Ramsbottom (1977) proposed further subdivisions of some of the "Major Cycles" but still related to eustatically controlled transgressive and regressive phases of deposition and of chronostratigraphic significance. He suggested the use of the term "Mesothem" first used by Hedberg (1973) in place of "Major Cycle" as part of a hierarchical classification of eustatically controlled, unconformity bounded units: Synthem, Mesothem and Cyclothem (Ramsbottom, 1977, p.281 and fig.9). Thus the Carboniferous in representing a major transgressive/regressive event forms a synthem which is built of a number of mesothems, approximately thirty in Britain (Ramsbottom, 1979). The mesothems in turn may be constructed from many minor cycles which are referred to the well established term cyclothem.

In an effort to standardise Dinantian stratigraphy in Britain, and to negate many of the regional problems of biostratigraphic correlation George et al (1976) established a series of regional stages (Fig.2). These were intended as definitive chronostratigraphic subdivisions based on precisely located stratotypes against which correlation was to be achieved "by any convenient means". In effect this implies biostratigraphic correlation since the stages are distinguished by characteristic faunal assemblages. They represent, therefore, merely the sumation of our understanding of Dinantian biostratigraphy to date ; our understanding of those "regional amendments" to the Vaughanian and other zonal schemes. The value of the stages rests with the positioning of the stratotypes within thicker sequences with reduced non-sequence.

The stages were said to "approximately coincide" with the Major Cycles of Ramsbottom (1973) "for each major transgression was accompanied by the migratory faunas that are used to recognise the different stages".

(c) Eustacy v. Tectonism

Since its first airing the concept of eustatically controlled cycles providing a basis for widespread correlation in the British Dinantian has been resisted. Few would dispute that Lower Carboniferous sequences throughout the world represent a major transgressive eustatic event, yet the great and often rapid lateral variations in thickness and facies, observed in Britain, has fostered the idea that tectonism was the dominant factor in controlling local cuvettes of deposition. That the effects of individual eustatic pulses can be recognised against this background of provincial diastrophism is thought, by many, untenable. A view recently brought into clear focus by George (1978). He has demonstrated the divergence of the regional stage boundaries from the mesothemic divisions of Ramsbottom (1973 and 1977) both in fact and theoretical concept. The former are designed to provide an objective and stable chronostratigraphic (in practice biostratigraphic) framework independent of variations in lithofacies or thickness. The mesothem boundaries on the other hand record marine transgressions and form the basis for an event stratigraphy, necessarily subjective in its application and requiring frequent amendment as our sedimentological understanding of Dinantian sequences improves. Contrast for instance the 'Major Cycles' of Ramsbottom (1973) with his 'Mesothems' of 1977.

Evidence for the eustatic origin of these sedimentary cycles rests on their widespread recognition and correlation. Much circular reasoning is involved here, however, since eustatic control is now so firmly entrenched in Dinantian thinking that it is often blindly assumed. The 'correct number' of cycles are 'recognised' and then become further supportive evidence for the eustacy hypothesis. However, as George (1979) points out in many British Dinantian sequences the objective evidence for such correlation is often either lacking, ambiguous or suggestive of more complex interpretation. The cycles of sedimentation which Ramsbottom has recognised are likely to result from an interplay of several factors eustatic, and tectonic, but also the much neglected yet potentially important effects of simple progradation. Any one of these factors may dominate over the others within particular cuvettes of deposition and with time. Thus, as Leeder (in discussion on George, 1979) suggests, a progadational shallowing in one area may, due to the imprecision of Dinantian biostratigraphy, be correlated with a tectonically induced regression elsewhere. The two events then misinterpreted as due to a single eustatic pulse.

These conceptual problems show that a cyclic or event stratigraphy is neither a replacement for, nor can operate without, the more traditional zonal stratigraphy. This stratigraphic debate, however, has tended to distract from the main value of Ramsbottom's work since there is little doubt that his event approach is potentially the more rewarding in terms of sedimentological and palaeoecological interpretation of local Dinantian successions. In either case we are in danger of putting the cart before the horse. First we must establish our correlations using all the techniques at our disposal, sedimentological, palaeontological and geological mapping. This done we are in a position to assess whether widely correlatable events of eustatic style do occur and the relative importance of local tectonism and progradation.

1.4 GEOLOGICAL SETTING

(a) Gross Palaeogeography

During the Dinantian the British Isles, excluding those parts of south-west England accreted during the Hercynian orogeny, formed part of the southern edge of the super continent of Laurussia (Scotese et al, 1979). Although moving steadily northwards in response to plate tectonic forces this southerly part of the huge craton lay within the equatorial belt for most of the system with the equator possibly passing through Anglesey even as the rocks considered herein were being laid down (Faller and Briden, 1978, fig.8).

Cratonic though its gross setting the British Dinantian records a period of great instability with rapidly subsiding troughs separating more ridged, often fault bounded, blocks (Fig.3). This differentiation formed in response partly to isostatically buoyant granite plutons e.g. Weardale Granite beneath the Alston Block (Dunham, Bott, Johnson and Hodge, 1961), partly to rejuvenation of underlying tectonic structures (George 1958) both of Caledonian ancestry. The story of the British Dinantian is one of the progressive, though far from regular, inundation of this complex archipelago as part of a world wide transgressive event.

In general the thicker sequences deposited in the troughs were of deeper water facies, dominantly mudstones and turbidites, whilst on the blocks shallow water carbonate shelf deposits accumulated. This largely marine phase of deposition was terminated by the southerly encroachment of Namurian and Westphalian deltas which effected a return to terrestrial conditions. According to Ramsbottom (1977) this represents the regressive phase of the Carboniferous Synthem. Detailed plate tectonic models related to the British Dinantian are proffered by Floyd (1972), Leeder (1976) and summarised by Anderton et al (1979) and Leeder (1982).

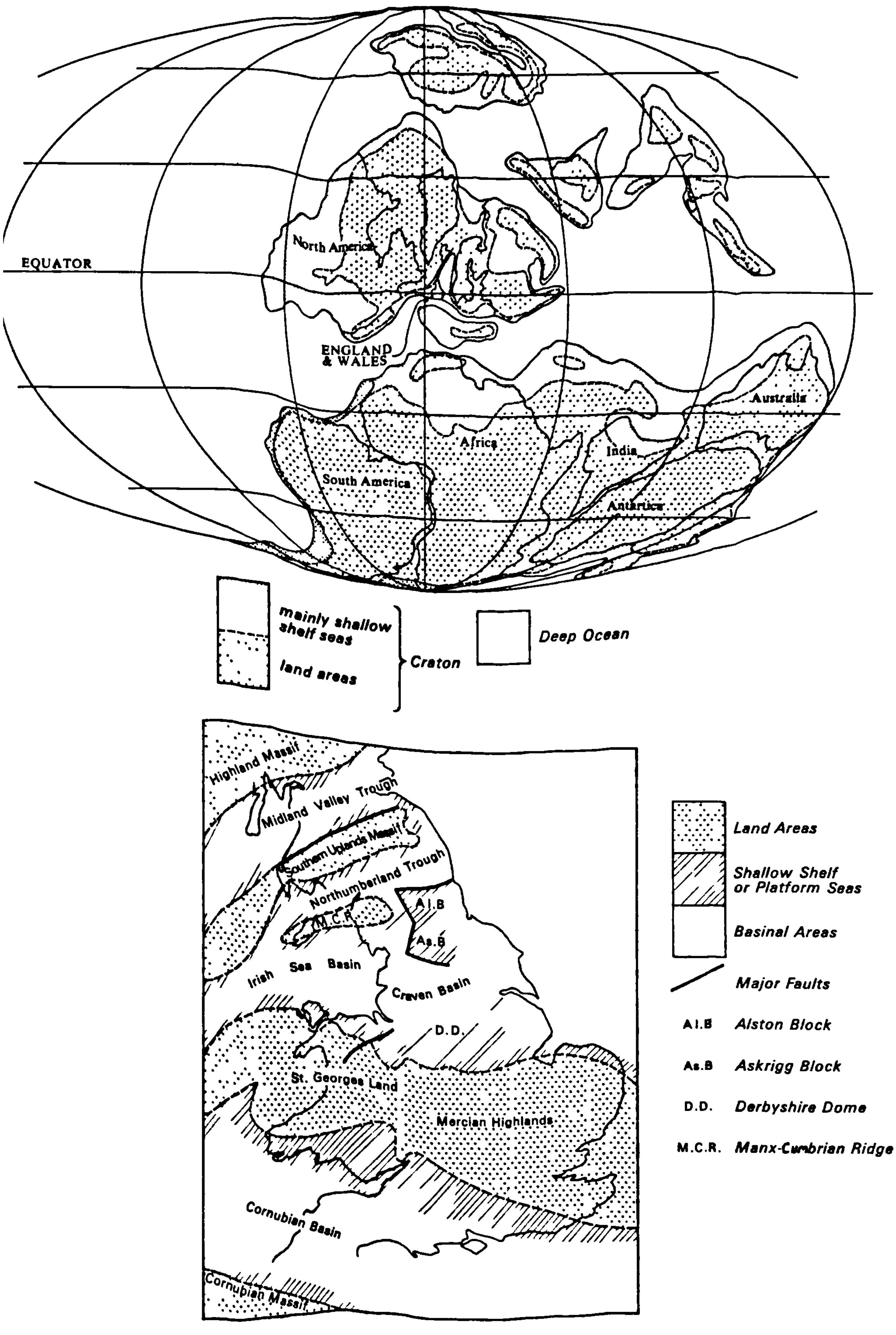


Fig.3 Dinantian Palaeogeography (after Scotese et al, 1979; Johnson, 1981 and George, 1958).

(b) Dinantian of North Wales

The Dinantian sequence in North Wales records the progressive drowning of the northern flank of St. Georges Land (Fig.3), the residual and deeply eroded core of a Caledonian upland, yet still tectonically active.

The Dinantian transgression reached this area late. Hind and Stobs (1906) showed that in North Wales only the highest of Vaughan's coral-brachiopod assemblages, the Dibunophyllum Zone could be demonstrated, but that both D_1 and D_2 sub-zones are present. These are now translated into the Asbian and Brigantian stages (George et al, 1976). To the unfossiliferous conglomerates and sandstones which occur sporadically at the base of the succession Hind and Stobs tentatively assigned a Seminula age (equivalent to the Holkerian), a zonal problem that remains unresolved.

Yet, whilst the sequence is attenuated in terms of time, it represents the thickest (up to 850 m) development of Asbian and Brigantian strata in the country (George, 1974). Contrast this with the Dinantian shelf succession in South Wales, of similar thickness (~ 1,000 m), but where all six stages are recognised and a tectonic differentiation between the northern and southern flanks of the St. Georges Land massif is demonstrated (op. cit. p.102). Nor in North Wales is this great thickness of strata evenly developed but occurs in restricted depressions which give way laterally to thinner sequences on adjacent highs. That this is in part an expression of the sub-Dinantian topography is undoubted, but George (1961) showed that differential subsidence was also active. He further demonstrated the coincidence of anticlinal and synclinal axes in the basement rocks to these highs and depressions respectively. He evoked, quite logically, the reactivation of these basement structures, yet to envisage the exact mechanism is more difficult. It seems unlikely that this was a renewal of compressive stress and was more probably an isostatic effect. Our understanding of these basement/cover relationships remains poor (but see Leeder, 1982).

This differentiation into depressions and highs was not of the same degree as the major trough and block development in so far as a deeper water basin was not created. Deposition within the depressions was always able to keep pace with subsidence and a shallow water carbonate shelf environment was sustained. The deeper water basin to which this shelf complex was marginal in fact lay to the north under what is now the Irish Sea, an extension of the Craven Basin, and here a full Dinantian sequence is thought to have accumulated (Ramsbottom, 1980).

The edge of the shelf was defined in part by a linear reef belt (Ramsbottom, 1969) remnants of which are seen at Dyserth and the Little Orme. But, although Mundy (1980) has shown that similar reefs on the northern margin of the Craven Basin occasionally develop a wave resistant framework at their crests, for the most part these were probably not part of a high energy shelf edge environment. Data presented by Neaverson (1929) and by Nutt and Smith (in prep.) suggests that the shelf edge was also the location for high energy shoal complexes, probably located slightly upslope from the reefs, and these protected the shallow, in parts restricted shelf lagoon. The shelf profile envisaged (Fig.4) whilst highly simplified, closely resembles the type I profile of Wilson (1975, p.361), a similarity strengthened by the cyclic nature of the lagoonal sequences (Somerville, 1979a, b, c and this thesis)

The recognition of this last feature has been of critical importance in our recent understanding of these North Wales sequences. Such sedimentary cyclicity is a characteristic feature of Asbian and

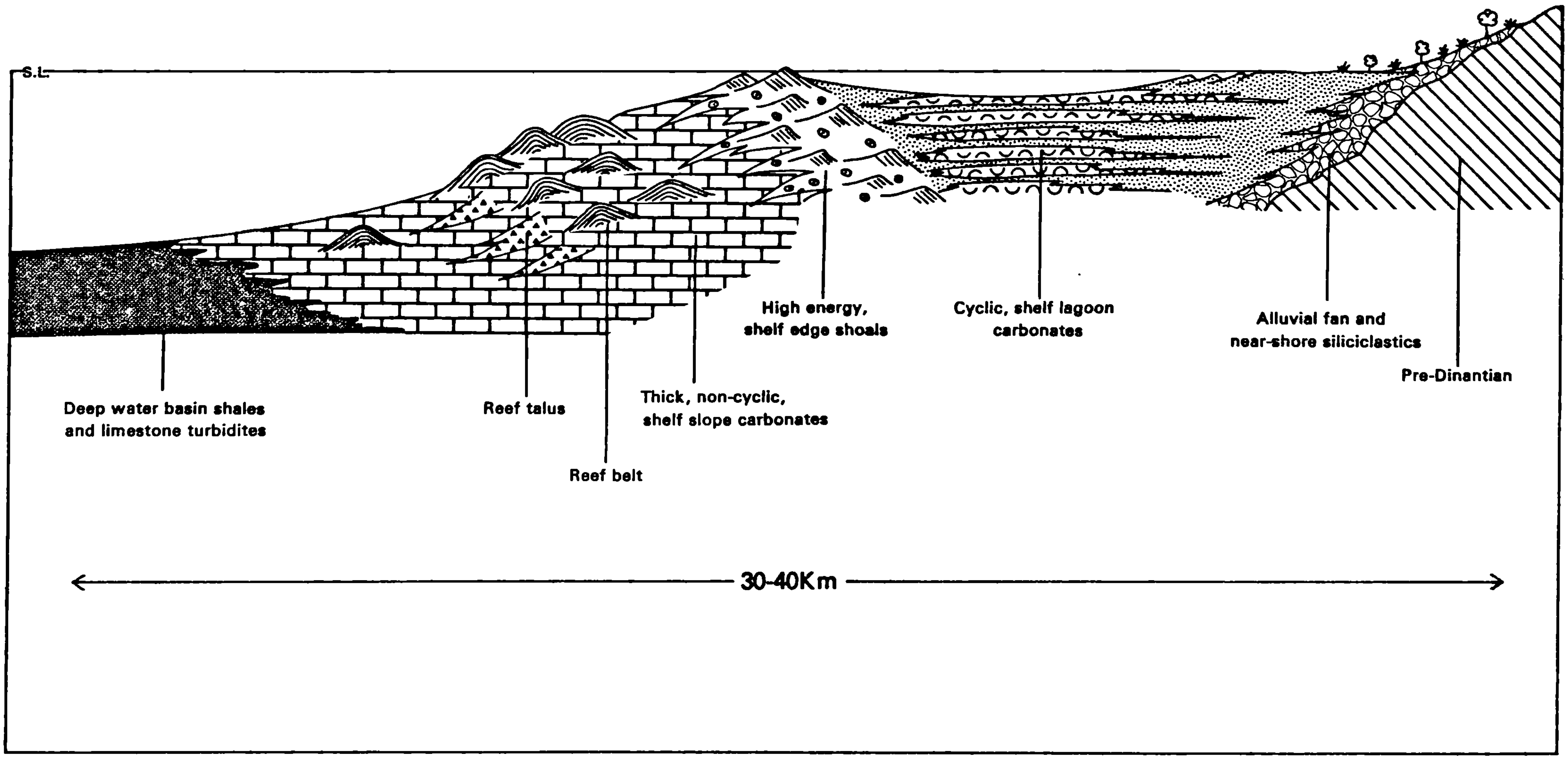


Fig.4 Suggested profile through the Dinantian shelf of North Wales

Brigantian shelf sequences throughout Britain and is referable to the minor cycles or cyclothems of Ramsbottom (1973, 1977).

(c) Dinantian of Anglesey

Within this regional picture the Anglesey Dinantian occupies a unique setting in representing the only significant thickness of strata preserved which accumulated on the landward side of the shelf lagoon. A sequence in which onlap of the Dinantian deposits over the older basement rocks is readily demonstrated and in which proximity of the shore was a major influence on sedimentation.

Early pioneer work on Anglesey by Henslow (1822) and later by Ramsay (1865) established the broad stratigraphy of the Island. The basal conglomerates and sandstones of the Lower Carboniferous were again a great source of confusion, initially, as elsewhere in North Wales, being assigned to the Old Red Sandstone.

Morton, who had previously established a crude lithostratigraphy based on colour for Dinantian sequences on the North Wales mainland (Morton, 1878, 1886, 1897, 1898), finally attempted to extend this to the Anglesey succession (Morton, 1901), but with only limited success (Section 3.2). It was Greenly (1919) in the Geological Survey Memoir to the Anglesey special sheet who first established some order to the succession. He was, like many of his day, preoccupied in his work on the Lower Carboniferous with applying the zonal scheme of Vaughan with consequent lack of attention to lithological detail. Nevertheless in recognising the D_1 and D_2 sub-zones and the contentious D_3, he was able to discover the broad sequence and structure of the three main areas of outcrop. He was thus able to demonstrate the gradual encroachment of Dinantian strata over the corrugated foundations of the ancient rocks of Anglesey. To show that deposition took place within two gulfs roughly corresponding to the Principal and Straitside Areas both trending SW-NE and opening to the north-east. Despite the Caledonoid trend Greenly resisted interpreting the gulfs as due to reactivation of basement structures, but preferred instead to envisage them as erosional hollows developed along the outcrops of the softer Ordovician and Devonian rocks which, of course, were of Caledonoid lineage.

More recent studies in the Penmon Area (Nicholls, 1962) and the Principal Area (Mittchel, 1964) have attempted to rectify Greenly's deficiencies in lithostratigraphy. These studies came during a period of increasing interest in limestone diagenesis, inspired by Bathurst's (1958) work on Dinantian limestones in North Wales, but in concentrating on this aspect of the lithologies paid scant attention to detailed environmental analysis on either a local or regional scale.

A fresh assessment of the palaeogeographical setting of the Anglesey sequences has been proffered in an eloquent paper by George (1974, p.108). He reinterprets much of Greenly's data and advocates a more dynamic depositional model. The gulfs he suggests were tectonically active, subsiding depressions separated by a positive Penmynydd Ridge (Fig.1) with some control by the Dinorwic and Berw Faults. Certainly it is these fracture zones, reactivated during Hercynian earth movements which largely determine the present outcrop pattern.

.5 METHODOLOGY

(a) Field Work

During the tenure of this research much emphasis has been placed on field observation achieved through geological mapping and the measurement of sections. Much of the Principal Area has been mapped at a 1:10,000 scale whilst the superb exposure along the north-eastern coast prompted mapping of this coastal strip at a 1:2,500 scale. Mapping of the Penmon and Straitside Areas has been less rigorous, a reflection of the poor inland exposure.

Measured sections at a 1:25 scale of the coastal cliff exposures in the Principal and Penmon Areas and to a lesser extent in the Straitside Area provide the main data base. These are augmented by measured sections of the major inland exposures both natural and quarry. The large scale of these sections, combined with the use of Dunham's limestone classification (see Section 5.2) has allowed the main bedding properties of the strata to be expressed graphically and in detail. These data are presented at the reduced scale of 1:50 in Charts 1 to 9.

Bed by bed sampling both for petrographic and palaeontological purposes was undertaken for all the accessible units and provided the basis for subsequent laboratory analysis.

(b) Laboratory Work

Limestone samples collected for petrographic analysis were cut and polished and the majority then either thin sectioned or acetate peeled. Many of the latter were also stained by mixed acidic solutions of Alizarin Red S and Potassium Ferricyanide using Dickson's Method (Allman and Lawrence, 1972). Quantatitive analysis of the various carbonate lithologies was achieved using a Swift point counter with automatic microscope stage attachment (for detailed discussion of techniques and references see Flugel, 1982). A limited number of limestone samples were analysized for insoluble residue content.

Palaeontological samples (mainly macrofauna) were cleaned and identified. Intermediate cutting and polishing was required for many of the corals. Etching of silicified fossils was also employed. No detailed statistical analysis of the physical parameters of Anglesey Dinantian faunas has been undertaken to date.

CHAPTER TWO

STRATIGRAPHIC FRAMEWORK

.1 INTRODUCTION

In common with Asbian and Brigantian strata on the North Wales mainland (Somerville, 1979a, b and c) and indeed throughout Britain (Ramsbottom, 1973) the Dinantian sequences on Anglesey are constructed from numerous minor transgressive/regressive cycles (or cyclothems; see Ramsbottom, 1977). In part these can be demonstrated by the ordered and predictable repetition of constituent lithofacies, but are recognised for the most part by the identification of palaeokarstic surfaces which punctuate the succession at frequent intervals. These are hummocky, sometimes pot-holed or channelled surfaces, the products of penecontemporaneous subaerial exposure of the limestone strata. They are characterised by the presence of rhizoliths, laminar micritic crusts and overlying bentonitic palaeosols. Palaeokarstic phenomena are discussed more fully in Chapter 3. Where channels incise through the surfaces these are filled by commonly complex siliciclastic sequences and these are described and interpreted in detail in Chapter 4.

Representing as they do relative falls in sea level, emergence and depositional diastems, palaeokarstic surfaces provided obvious criteria by which to subdivide the succession. Some of these horizons were recognised by previous workers, notably the sandstone piped limestones of Greenly (1919, p.613), but hitherto their number and significance has gone unheeded.

The recognition of marked minor cyclicity, coupled with the detailed mapping and measurement of sections, particularly in the coastal areas, has enabled the construction of a more precise local stratigraphy which supersedes those of previous workers (Figs. 5, 6 and 7).

.2 PRE-EXISTING CLASSIFICATIONS

In applying different criteria in an effort to subdivide the Anglesey Lower Carboniferous sequence, previous workers produced differing and often confusing accounts of the local stratigraphy.

Morton, in a series of papers (1878, 1886, 1897 and 1898) had, with some success, subdivided Lower Carboniferous sequences on the North Wales Mainland on the basis of colour. In a publication after his death (Morton, 1901) he tried to extend his three fold division of Lower Brown Limestone, Middle White Limestone and Upper Grey Limestone to the Anglesey sequence. His description and interpretation of the sections are at considerable variance with the succession demonstrated here and it is clear from the account that he did not recognise the cyclic nature of the succession or the degree of repetition by faulting. Paradoxically, however, once these facets of the sequence are taken into account broad divisions, comparable with those of Morton for the mainland outcrops, re-emerge (see Section 2.5).

Greenly (1919, p.608) established the coral-brachiopod zonal sequence for the area. He suggested an S_2 zone age for the lower positions of the Lligwy Sandstone (p.616) following established practice for the unfossiliferous basal sandstones in North Wales. He recognised with certainty the presence of D_1 and D_2 subzones, but further argued for the existence of the higher D_3 subzone (p.609).

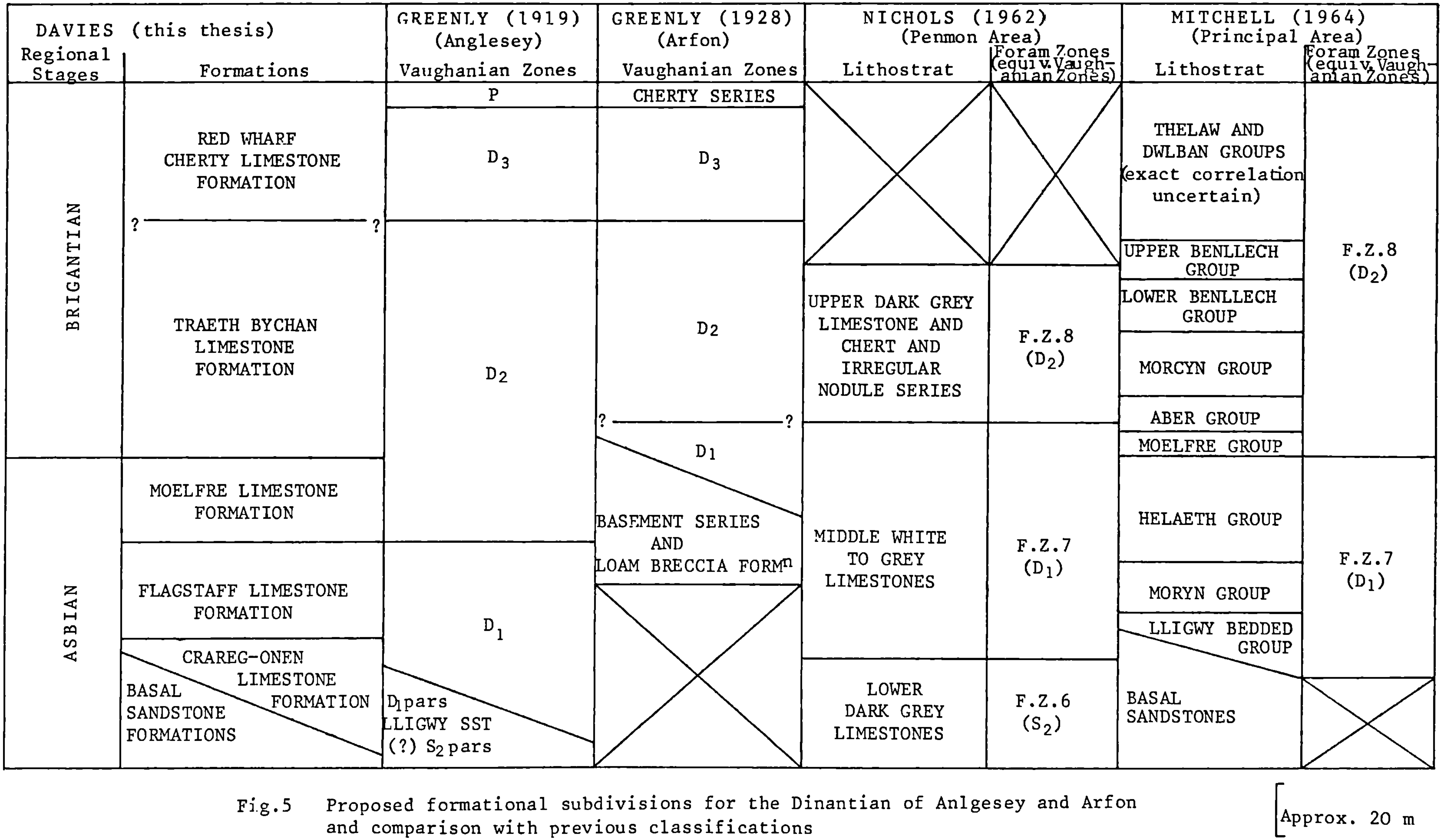


Fig.5 Proposed formational subdivisions for the Dinantian of Anlgesey and Arfon and comparison with previous classifications

The bedded cherts which cap the sequence in the Principal Area were, on further tenuous evidence assigned to the Posidonomya zone (Dixon and Vaugan, 1912) of topmost Dinantian age. An interpretation supported, Greenly thought, by unconformity at the base of the overlying Millstone Grit.

In its simplest and non-controversial form (i.e. distinguishing D_1 and D_2), Greenly's succession, whilst not coinciding with the subdivisions proposed in this thesis, does broadly parallel them, at least in the Principal and Straitside Areas. The existence of a D_3 subzone is now widely disputed (George, 1969). In North Wales, beds rich in Lonsdaleia duplicata (Mart.) were assigned a D_3 age (Neaverson, 1929), but this practice has been shown untenable by the discovery of typical D_2 goniatites from the same beds (Neaverson, 1943). The exact age of the cherts remains questionable.

There are also serious deficiencies in the detail of Greenly's account resulting from the scant attention he paid to lithologic sequence. This is most notable in the Penmon Area where his mis-correlation east to west is gross. He failed to recognise equivalent strata exposed at Careg-onen and around Penmon Priory. Indeed he seems to have been so preoccupied with a model for the structure and sequence in the area (p.622) as to disregard the profound degree of faulting and eventually to refute the faunal evidence he had otherwise relied upon so strongly. Thus in a footnote referring to Flagstaff Quarry on page 653 he notes that Prof. Garwood had "collected . . . Daviesiella llangollensis (Dav.) in the lower part of this quarry and is therefore inclined to suspect the existence of S_2. But the stratigraphy does not favour this." Greenly interpreted these beds as high in D_2 and therefore failed to recognise, as Garwood had suspected, the oldest of the limestone strata in the Dinantian sequence on Anglesey. Following on from this and again through his unwillingness to adopt a systematic lithostratigraphic procedure he was unable to correlate, other than in zonal terms, between the areas. This in turn raises serious doubts about his palaeogeographic interpretations.

Greenly's lithostratigraphic nomenclature is confined to local descriptive terms e.g. Lonsdaleia floriformis beds, Cliff-Brow limestone, junceum- limestones, etc.,lacking precise definition. He did, however, distinguish the major sandstone bodies on the map and his names for these are largely retained.

More recent studies have attempted to rectify Greenly's dificiencies in lithostratigraphy. Nichols (1962), as part of a regional study on the Dinantian of North Wales, erected a lithostratigraphy for the Penmon Area which was, in effect, a more sophisticated version of Morton's scheme, but with more rigorous lithologic definition. This was set against standard zonal schemes, but Nichols felt that macrofaunal zonation was too imprecise and resorted to a microfaunal study, recognising zones 6, 7 and 8 of Cumming's foraminiferal scheme. These he equated with coral-brachiopod subzones S_2, D_1 and D_2 respectively. The existence of the S_2 subzone was, at the time, given added credence by the occurrence of D.llangollensis in the same strata.

Mitchell (1964) undertook a similar study on the Principal Area again erecting a local lithostratigraphy in which he recognised foraminiferal zones 7 and 8 only. He felt there was "no positive evidence for the presence" of zone 6. No detailed lithostratigraphic correlation with Nichols' subdivisions in Penmon was attempted.

As part of a wider ranging study Nichols' lithostratigraphy was necessarily broad in its definition and he failed to recognise the minor cyclicity displayed by the sequence. Nevertheless the broad formational groupings of the minor cycles, outlined below (Section 3.4), do closely correspond with Nichols' subdivisions for the Penmon Area.

The existence of S_2 limestones in Anglesey and indeed throughout North Wales is now refuted primarily because of the redefinition of the zonal assemblages (George et al, 1976). Thus D.llangollensis is now regarded as a characteristic low Asbian (D_1) form. It would follow that both foraminiferal zones 6 and 7 now correspond with D_1. Unfortunately Cummings' foraminiferal scheme was never published in full and therefore its equivalence with the Vaughanian coral-brachiopod scheme was not widely tested. It is certainly open to similar criticism of strong facies control of the zonal assemblages. Thus it is not clear whether D_1 can indeed be further subdivided into two foraminiferal zones, or whether the lower zone 6, coinciding as it does with an unusual lithofacies development (Section 3.4b) is facies dependent and of local significance only.

Mitchell's stratigraphic subdivision, dealing only with the Principal Area, was more detailed than that of Nichols but his account is far from lucid. He did however recognise a broad alternation of lithologies in which are the seeds of minor cyclicity which forms the basis of the present study. His subdivisions were initially based on objective lithostratigraphic criteria with like strata grouped together in what might otherwise be regarded as a quite logical arrangement. In fact for practical reasons of mapping he was often forced to combine lithologically dissimilar units and this subtracts from its value. More seriously his scheme is now harder to accept because the palaeokarstic surfaces which form the basis of this present study occur within his divisions. His "groups" are in fact of variable status relative to the minor cycles with some corresponding to a part of a single cycle and others embracing several.

Mitchell rightly and prophetically questioned the presence of foraminiferal zone 6 in the Principal Area, but in restricting his work to this area he failed to realise the full significance of his results. These now provide welcome supporting evidence for the stratigraphic distribution outlined below bearing in mind the probable facies control mentioned above.

The D_1 (≡ foram. zone 7) / D_2 (≡ foram. zone 8) boundary proffered by Nichols and Mitchell in their respective areas is discussed more fully in Section 3.6 as this zonal boundary is now thought to be equivalent to the Asbian/Brigantian stage boundary (George et al, 1976).

2.3 CYCLIC STRATIGRAPHY

These previous accounts, both individually and collectively lacked a unifying theme to enable the more simple description and interpretation of the succession. Minor cyclicity provides such a theme and one, moreover, ideally suited to facies analysis and environmental interpretation.

The conceptual basis of cyclic stratigraphy is discussed more fully in Section 5.6 which deals with the sedimentological interpretation of the minor cycles. In essence it represents event stratigraphy and clearly, from a theoretical standpoint, it is important to recognise what sort of events a particular form of cyclicity records as to whether it provides a meaningful basis for a stratigraphic framework. In practice, however, stratigraphies based on cyclicity are erected because: (1) cyclicity is readily recognisable; (2) it can be used to form a workable local stratigraphic framework;

and (3) it then allows the ready interpretation of the sequence in terms of depositional events. Thus in the Anglesey Dinantian: (1) the frequent occurrence of palaeokarstic surfaces allows the recognition of minor cyclicity; (2) correlation of palaeokarstic surfaces achieved by standard stratigraphic procedure i.e. using marker horizons, faunal content, thickness considerations etc., provides a stratigraphic framework; and (3) interpretation of the palaeokarstic surfaces as recording periods of emergence allows the interpretation of the minor cycles, in the simplest sense, as resulting from alternating transgressive and regressive events.

In the Anglesey Dinantian 26 minor cycles (excluding those in the Careg-onen Limestone Formation, see Section 4b(i), in which there are at least a further 4 minor cycles) have been recognised in rocks which span the Asbian and Brigantian stages. These stages span approximately 15 million years between them (George et al, 1976 p.76 and Ramsbottom, 1979) and so each minor cycle and its accompanying emergent episode represents on average 500,000 years (clearly much of this time is likely to be taken up by the emergent episode and is represented in the palaeokarstic surfaces). The cyclic stratigraphy outlined below, therefore, approximates, at least locally, to a chronostratigraphic framework with each palaeokarstic surface being regarded as essentially isochronous (Wilson, 1975 p.53). The minor cycles provide a means of subdivision far beyond the resolution of current Dinantian biostratigraphy (Ramsbottom, 1979).

Ramsbottom (1977) advocated an eustatic and therefore widespread chronostratigraphic significance to his mesothemic divisions and, by implication, to their constituent cyclothems. Whether the minor cyclicity recognised in the Anglesey Dinantian is truly cyclothemic in character, using Ramsbottom's definition, and provides a basis for wider regional correlation, can only be ascertained by the painstaking tracing of individual minor cycles between adjacent sections, rather than by glib comparison of widely separated areas (see Section 2.5).

There are, however, problems inherent in relying too strongly on minor cyclicity as a basis for stratigraphic subdivision. Its recognition is at times subjective and there are practical difficulties in tracing individual minor cycles in areas of poor exposure and in faulted ground. To resolve these difficulties, to facilitate description and to aid the regional correlation the minor cycles have been grouped together into broader stratigraphic units.

Grouping of minor cycles into more managable packages may be achieved in three main ways: (1) a lithostratigraphic approach, grouping a run of lithologically similar cycles i.e. cycles which exhibit a similar pattern of constituent lithologies; (2) by picking particularly marked palaeokarstic surfaces in representing pronounced non-sequence; and (3) by palaeontological zonation.

Macro and to a lesser extent micropalaeonotological zonation have allowed the recognition of the Asbian and Brigantian stages. These subdivisions, however, are too broad to be of practical use and fail to describe the patterns of minor cyclicity evident in the succession.

With the palaeokarstic method in keeping with the hierarchical procedure of Ramsbottom (1977), such broader groupings of minor cycles would represent mesothems. Mesothems, however, are only capable of definition within conformable sequences (Holland et al, 1978 p.15). Where there is significant non-sequence as in Dinantian shelf sequences their recognition rests with an assessment of the relative importance of palaeokarstic surfaces. Certain palaeokarstic surfaces, which appear to represent particularly marked regressive events, do occur within the Anglesey Dinantian succession and could be used to define mesothems. Indeed Ramsbottom (1977 p.283) cites evidence presented by George et al (1976) from the Anglesey sequence for subdividing the Asbian into his mesothems D5a and D5b (Fig.2). There are, however, problems in assessing relative maturity of palaeokarstic surfaces. Palaeokarstic phenomena tend to be variably developed along individual surfaces with deep "mature" palaeokarstic profiles giving way laterally to unaltered host rock with perhaps a thin laminated micritic crust and a few rootlets which might reasonably be regarded as "immature". Nor is the occurrence of deep sandstone-filled channels incised through palaeokarstic surfaces necessarily indicative of a particularly pronounced fall in base level. The channels have a shoestring geometry and their outcrop is largely fortuitous. Most, if not all, palaeokarstic surfaces may have such features associated with them. Further research, in particular correlation with more complete basinal sequences, may show some of the apparently more marked palaeokarstic surfaces to be of widespread significance and to be truly mesothemic in rank.

For the moment, however, the emphasis in grouping minor cycles in to broader divisions has been placed on lithostratigraphic criteria and they are accordingly termed formations. It should be stressed, however, that the boundaries of these formations are defined not at points where characteristic lithologies appear or disappear but at the nearest convenient palaeokarstic surface i.e. the formations comprise whole numbers of minor cycles. The formations are not in any sense arbitrary, however, but take account of natural groupings of the minor cycles by way of similar patterns of constituent lithofacies and/or the occurrence of some distinctive lithology. Such formational groupings are undoubtedly present and raise the important question of why they should occur at all. Why should several consecutive minor cylces each resulting from a separate transgression onto the Anglesey Dinantian shelf exhibit such similarities? These formational groupings are evident across much of the North Wales crop (Section 2.5). Their origin and implications are considered in more detail in Section 5.7.

2.4 STRATIGRAPHIC NOMENCLATURE

Following standard lithostratigraphic procedure the formations are named after the location of their type sections. The minor cycles provide more of a problem. The finer subdivisions of formations are, strictly speaking, members (Holland et al, 1978) yet the minor cycles are themselves composed of members. To negate these problems many authors have numbered minor cycles (e.g. Somerville, 1979a, b and c), whilst a long established practice in the Yoredale type cycles of Northern England has been to name the conspicuous limestone member after its type locality, this name then acting as a label for the cycle as a whole e.g. Burgess and Mitchell (1976).

The use of names rather than numbers is preferred since it provides greater flexibility. If subsequent to this study further palaeokarstic surfaces are recognised within a minor cycle, allowing its further subdivision, named cycles may be easily split into upper, middle and lower. In a numbered scheme all overlying cycles must be renumbered or else a cumbersome dual numbering system must be introduced (e.g. 2a, 2b, etc.). A further and more serious drawback of the numbering scheme is that it may lead to careless comparison between areas, prompting correlation by numbers without detailed attention being paid to

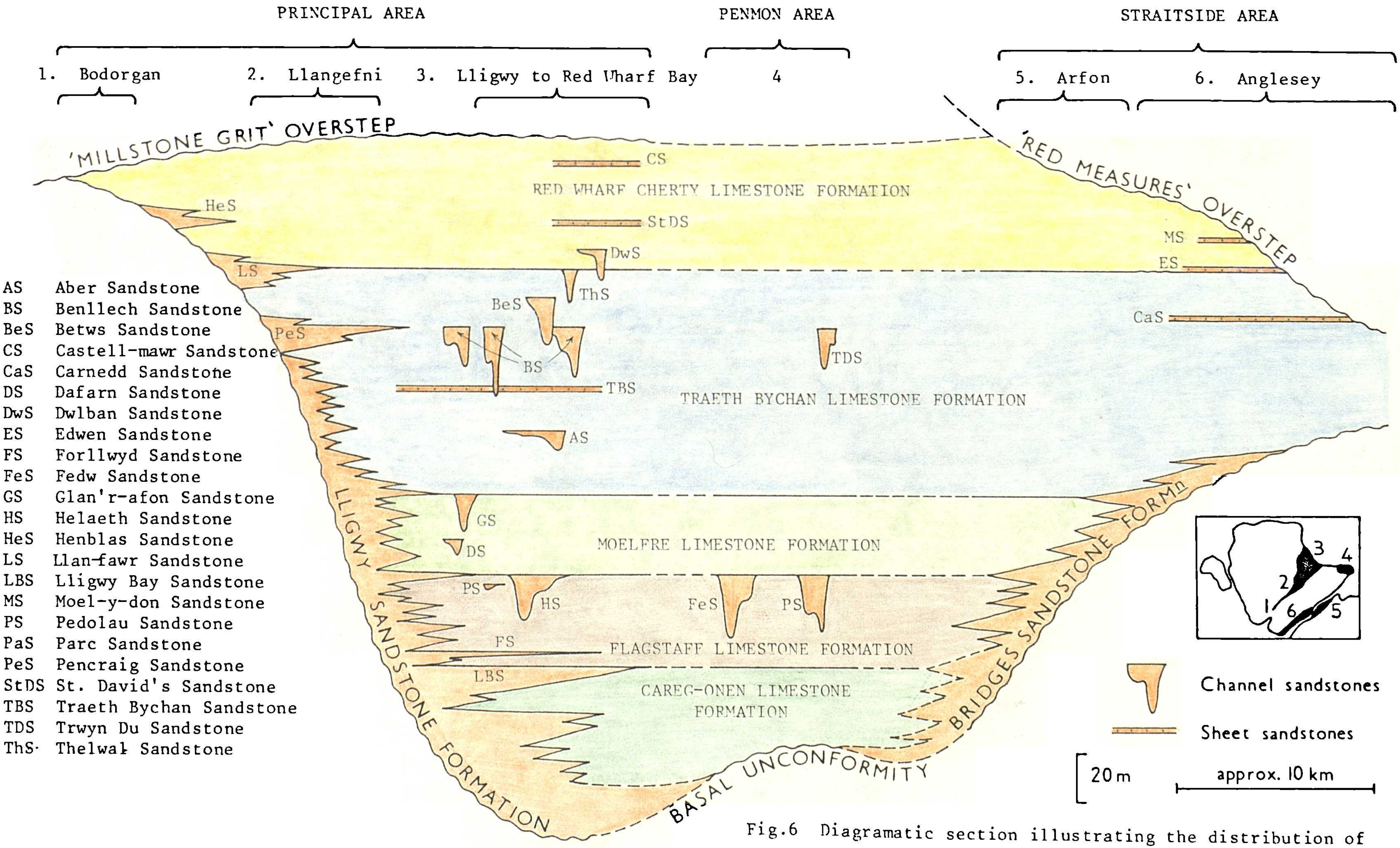


Fig.6 Diagramatic section illustrating the distribution of sandstones and the overlap of formations in the Dinantian of Anglesey and Arfon.

Regional Stages	Formns	MINOR CYCLES: PRINCIPAL AREA	PENMON AREA	STRAITSIDE AREA
BRIGANTIAN	RED WHARF CHERTY LIMESTONE FORMATION	Castell-mawr Beds Castell-mawr Sst		Red Wharf Cherty Limestone Formation undivided
		St. David's Beds St. David's Sst		
		Upper Dwlban Beds		
		Dwlban Sst		
		Lower Dwlban Beds ?		? ?
	TRAETH BYCHAN LIMESTONE FORMATION	Thelwal Sst		The formation has been sub-divided into the following minor cycles: Spring Beds Upper Carnedd Beds & Carnedd Sandstone Lower Carnedd Beds Upper Dyke Beds Lower Dyke Beds but correlation with other areas is not yet possible.
		Pen-y-coed Beds		
		Betws Sst		
		Upper Dinas Beds	TB7	
		Beblech Sst	Trwyn Du Sst	
		Lower Dinas Beds	TB6	
		Upper Morcyn Beds Traeth Bychan Sst	TB5	
		Lower Morcyn Beds	TB4	
		Porth-y-Rhos Beds	TB3	
		Aber Sst Porth-yr-Aber Beds	TB2	
		Eglwys Siglen Beds	TB1	? ?
ASBIAN	MOELFRE LIMESTONE FORMATION	Glan'r-afon Sst Upper Harbour Beds	M8	Moelfre Limestone Formation undivided
		Lower Harbour Beds	M7	
		Upper Lookout Beds	M6	
		Lower Lookout Beds	M4 and M5	
		Dafarn Sst Upper Helaeth Beds	M3	
		Middle Helaeth Beds	M2	
		Lower Helaeth Beds	M1	BRIDGES SANDSTONE FORMATION
	FLAGSTAFF LIMESTONE FORMATION	Helaeth Sst	Fedw & Parc Sst	
		Royal Charter Beds	F6	
		Pedolau Sst Pedolau Beds	F5	
		Moryn Beds	F4	
		Porth Forllwyd Sst	F3	
		Forllwyd Sst Lligwy Beds	F1 and F2	
	CAREG-ONEN LIMESTONE FORMATION / BASAL SANDSTONE FORMns	Lligwy Bay Sst Limestones at Careg-ddafad LLIGWY SANDSTONE FORMATION	Careg-onen Limestone Formation undivided ? ?	

Fig.7 Stratigraphic nomenclature and minor cycle correlation for the Dinantian of Angelsey and Arfon (no thicknesses implied)

Channel Ssts

Sheet Ssts

lithological sequence and palaeontology.

It is however the nature of the sections which largely determines which scheme is used. Thus in the Principal Area the extended coastal crop of the minor cycles has allowed their individual naming after suitable localities, whilst in Penmon the precipitous cliff and quarry sections make a system of numbering more appropriate. For the present both schemes are retained and correlation between them shown in Fig.7. Ultimately it is hoped that, as with the Yoredale limestones, the names from the Principal Area will usurp the use of numbers in Penmon, and when better correlation has been achieved also replace the separate set of names used in the Straitside Area.

Unfortunately the minor cycles contain no conspicuous repetative lithology like the limestone members of the Yoredales, e.g. Hardraw Limestone, Jew Limestone etc. During the early stages of field work the cycles were termed "Beds" e.g. Pedolau Beds, Porth-yr-Aber Beds etc. and this terminology has been retained (Arguably the use of the term cycle would be better e.g. Pedolau Cycle, Porth-yr-Aber Cycle etc.).

The succession has been subdivided into the following formations and minor cycles (see Figs.5, 6 and 7).

(a) Basal Formations

In the Principal and Straitside Areas basal siliciclastic formations separate the overlying limestone sequences from the older basement rocks of the Island (Figs.8 and 9).

In the Principal Area the Lligwy Sandstone Formation (after Greenly, 1919 p.616) forms a diachronous wedge up to 45 m thick at Lligwy Bay, thinning southwestwards as the Dinantian succession onlaps across Devonian, Ordovician and Precambrian strata.

A similarly diachronous but lithologically more varied basal

FIG. 8. SKETCH MAP OF THE
PRINCIPAL AREA
(partly after Greenly, 1919)

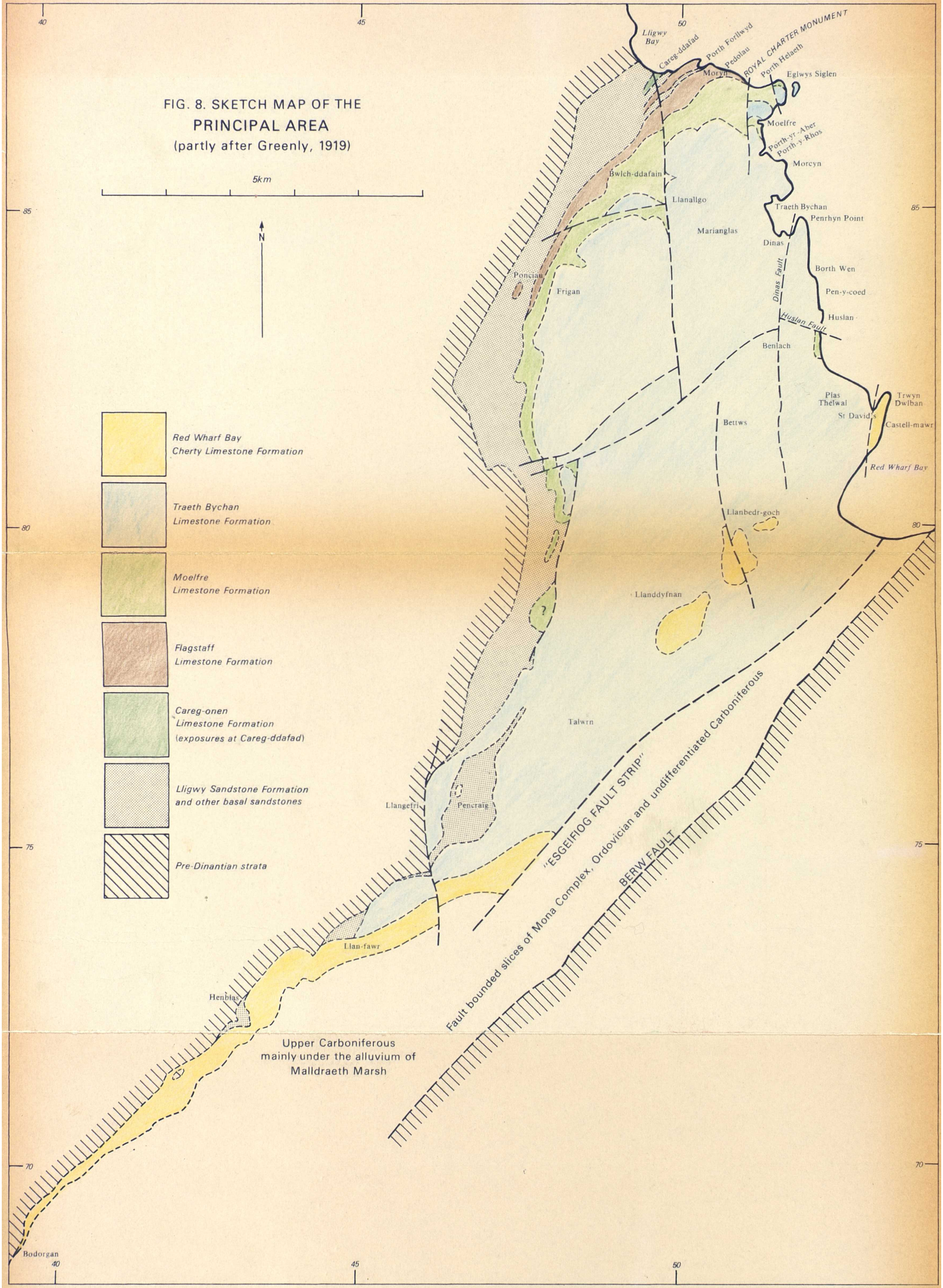
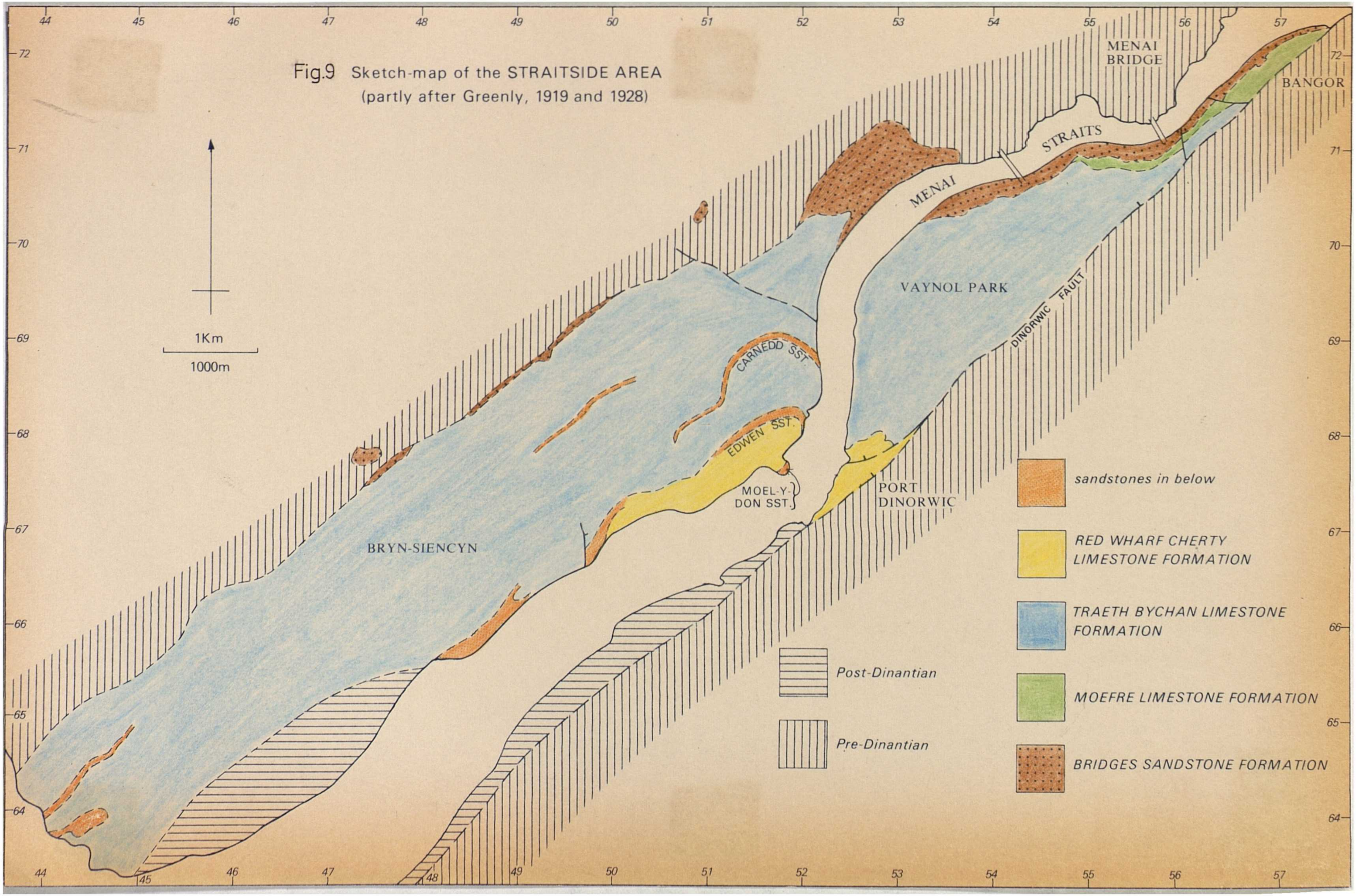


Fig.9 Sketch-map of the STRAITSIDE AREA (partly after Greenly, 1919 and 1928)



siliciclastic formation occurs in the Straitside Area. Units of this deposit on the Arvon side of the Straits were described by Greenly (1928) as his "Basement Series" which apart from conglomerates and sandstones also included his interesting "Loam-Breccia Formation" (op. cit. p.386). The Fanogle Sandstone of Greenly (1919 p.619) represents the equivalent basal sequence on Anglesey. The stratigraphic nomenclature for these basal deposits can now be rationalised and a new name the Bridges Sandstone Formation is proposed. This reflects the location of the best sections through the formation, including its unconformable contact with Ordovician slates, between the Menai Suspension and Britannia Railway Bridges on the mainland side of the Straits.

No basal terrigenoclastics are exposed in the Penmon Area although Nichols (1962) records conversations with local quarry men who reported sandstones beneath the lowest limestones at Tan-dinas Quarry.

(b) Careg-onen Limestone Formation

This is the oldest of the limestone formations in the Anglesey Dinantian. It is only recognised with certainty in the Penmon Area although purely on thickness considerations the limestones below the Lligwy Bay Conglomerate may be part of laterally equivalent units in the Principal Area.

(i) Penmon Area (Fig.10, Chart 1)

The formation is best seen and most fully developed around Careg-onen, in Tan-dinas Quarry and the cliffs to the east, at the western end of the Penmon Area. This has been chosen as the type section. Upper parts of the formation are also well observed in Flagstaff Quarry to the east and further small quarry

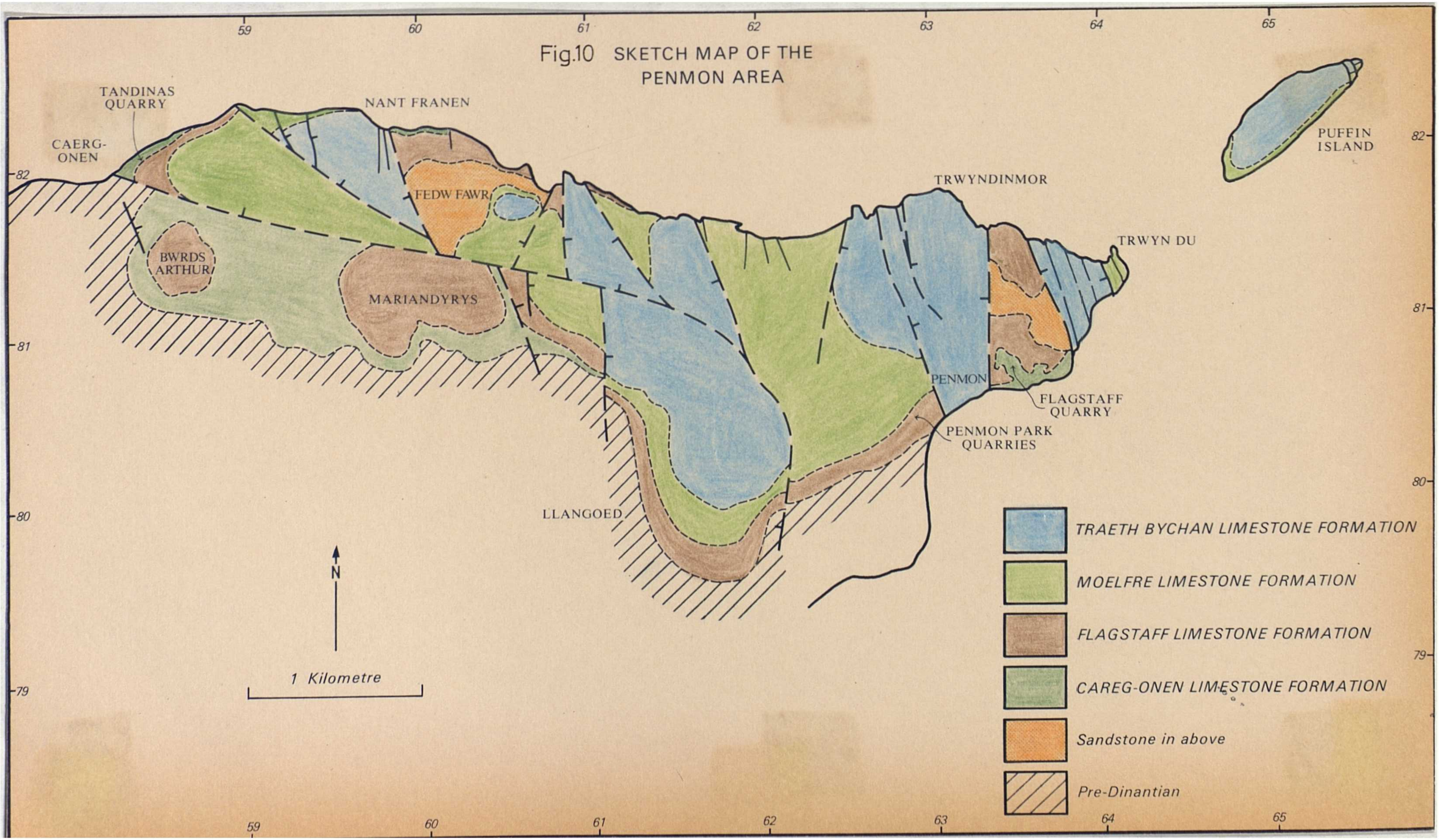


Fig.10 SKETCH MAP OF THE PENMON AREA

sections occur along the inland escarpments in the south of the Penmon Area.

At Careg-onen the formation attains its greatest observed thickness of 41 m. The base of the formation is not seen, but as discussed above, it is thought to rest on (?) basal sandstones formerly seen during quarrying. Fallen blocks now conceal the lowest 5 to 6 m of the quarry face. Assuming that the lowest couple of metres are of sandstone, the maximum thickness obtained for the Careg-onen Formation is 45 m.

The formation is distinctive in showing a considerable development of white weathering, dark grey to fawn, porcellanous calcite mudstones (micrites) which commonly exhibit well formed birdseye structures (Ham, 1952; Shinn, 1968 and Deelman, 1972). Rootlets and occasional stromatolitic lamination also occur in this lithology. At Careg-onen, above a lower run of dark, argillaceous, skeletal packstones, the formation is predominantly composed of these birdseye micrites. They occur in beds of varying thickness up to 1.50 m and are nearly always separated by bands or occasionally quite thick beds of dark grey to black, carbonaceous shale. This interbedding imparts a rhythmic appearance to the succession. Lithological rhythmicity is further observed towards the middle of the formation where several individual beds exhibit upwards fining each with coarse, often oncolitic, skeletal grainstone at their base, grading through finer skeletal grainstone and packstone into the capping calcite mudstone phase. Discrete beds of skeletal grainstone, often cross bedded and with lenses of oncolitic and skeletal coquina also occur.

At Flagstaff Quarry calcite mudstones are again conspicuous but less dominant. Here the lowest beds exposed broadly equate (see Chart 1) with the fining upwards rhythms described from Tan-dinas Quarry, but here they have given way to thick units, up to 8 m, of cross bedded skeletal grainstone, oolitic in parts, with the calcite mudstone phase occluded. A further 5 m thick skeletal grainstone unit with oncolitic and coquinoid lenses tops the formation in contrast to the calcite mudstones in the type section. The two lithologies are clearly close lateral equivalents of one another. This lateral variation precludes the correlation of individual beds between the two sections.

Inland outcrops and features to the south of Bwrdd Arthur [SH 5850 8125] suggest that the formation has thinned to 30 m and indicate that overlap is taking place southwards and possibly westwards against a gently rising floor of Ordovician rocks. The identification of the formation in the east of the area at Flagstaff Quarry precludes any significant overlap in this direction as was proposed by Greenly (1919 p.626) and since George (1974) used Greenly's data his palaeogeographic reconstruction is similarly erroneous (see also Section 2.7).

The top of the formation is defined by a well developed palaeokarstic surface formed on the uppermost calcite mudstone bed at Careg-onen. This is a markedly hummocky surface and is overlain by a conspicuous red and grey mottled mudstone palaeosol. At Flagstaff Quarry the surface is equally marked but here it is developed on skeletal grainstone and has associated laminated micritic crust, rhizoliths and overlying bentonitic clay. Despite these differences in host lithology, marker horizons in the overlying Flagstaff Limestone Formation leave no doubt that this is indeed the same palaeokarstic surface in both localities.

In contrast with overlying formations, palaeokarstic surfaces are not readily identified within the Careg-onen Formation. This is somewhat surprising since the calcite mudstones which predominate are thought to record deposition on emergent, supratidal flats (see Section 5.5) prone, one might have thought, to karstification with any minor fall in base level. Palaeokarstic surfaces do occur and so therefore minor cyclicity can be said to be present, but how does this relate to the lithological rhythmicity also observed within the formation ? Can this be generated by purely sedimentary mechanisms or does it too reflect a base level control? At what stage does an emergent carbonate mud flat with birdseye structures and rootlets become karst? These questions are discussed more fully in Chapter 5 Section 6 , but because of the problems they pose and also due to the rapid lateral variations in facies, formal subdivision of the formation into minor cycles has been avoided.

(ii) Principal Area

In the Principal Area deeply eroded limestones below the Lligwy Bay Conglomerate at Careg Ddafad [4979 8712] (see Section 4.7) are, on thickness considerations below marker horizons in overlying formations, likely to be laterally equivalent to the Careg-onen Limestone Formation. Thickness variation between the areas for overlying formations is not pronounced and therefore equivalence based on such criteria seems justified.

These units have been described by Cope (1975) who observed the outcrop after natural shifting of beach sands provided a more extensive section than is now available. He recognised the insitu nature of these limestones, which contain shelly lenses, and the erosive contact of the overlying Lligwy Bay Conglomerate.

This conglomerate and the succeeding finer sandstones and shales form, on the coast at least, a distinct and higher clastic sequence than the Lligwy Sandstone Formation proper. It is clear from inland mapping however that the limestones pass rapidly into the basal sandstone formation and that the Lligwy Bay Sandstone and Conglomerate are really extensions of it (Fig.11).

Detailed correlation with the Penmon Area is, of course, precluded. No calcite mudstones so characteristic of the formation are exposed. The limestones with shelly lenses are reminiscent of the skeletal grainstone lithologies in Penmon, but are not exclusive to the Careg-onen Formation.

An attractive aspect of such a correlation is that the Lligwy Bay Conglomerate and Sandstone which rest on these limestones would then represent a major clastic influx at the emergent phase which marks the top of the formation. Overlying limestone strata are readily correlatable with the lower parts of the succeeding Flagstaff Limestone Formation in the Penmon Area.

(iii) Age of the Careg-onen Limestone Formation

The occurrence of *Dibunophyllum* sp. towards the top of the formation in Tan-dinas Quarry firmly links it with the Vaughanian zone of that name, but it is the occurrence of *Daviesiella llangollenis*, often in abundance, throughout these strata which is of greatest stratigraphic significance. As discussed above this form has previously been thought indicative of an S_2 age and quite recently its common occurrence prompted Cope (1975) to advocate such an age for these particular rocks. The problem was that whilst *D.llangollensis* had been recognised as of stratigraphic value in Northern England and North Wales it had not, until very recently

(Ramsbottom, in discussion; George, 1978 p.255) been found in the Bristol and South Wales areas where the original zonal assemblages were defined. Recent finds in South Wales confirm the conclusions already reached by George et al (1976) that D.llangollensis is in fact an Asbian (equivalent to D_1) form and moreover is characteristic of the lower part of that stage. The Careg-onen Limestone Formation is accordingly regarded as of early Asbian age.

The formation embraces the whole of Nichols' Penmon Lower Dark Grey Limestones which he argued belonged to Cummings' Foram Zone 6. We have discussed the likely facies control of this zonal assemblage as it coincides with an unusual predominance of calcite mudstones. The occurrence of D.llangollensis throughout North Wales and Northern England is similarly linked to these lithologies (Burgess and Mitchell, 1976) and is likely to be no less facies dependent. Its status as a diagnostic zonal form should therefore be viewed with caution.

Cope (1975) recognised a D_1 fauna from the limestones below the Lligwy Bay Conglomerate in the Principal Area.

(c) Flagstaff Limestone Formation

The Flagstaff Limestone Formation succeeds the Careg-onen Limestone Formation in the Penmon Area and is the lowest of the limestone formations recognised with precision in the Principal Area. The base of the formation is only readily identified in the former area and it is here necessarily that the type section must be located.

(i) Penmon Area (Fig.10, Chart 2)

The formation is well seen overlying the Careg-onen Limestone Formation in Tan-dinas Quarry, but a fuller sequence is observed in Flagstaff Quarry, with upper parts of the formation occurring in the adjacent Penmon Park Quarries. This has been chosen as the type area. The formation has also been recognised in the cliffs north of Fedw-fawr. Further inland outcrops have not been examined in detail.

Since the complete formation is never seen in full in any one locality the thickness of 37 m is a compilation from several localities. This precludes any attempt at estimating thickness variation for the formation as a whole across the Penmon Area. Thickness variation within individual minor cycles is minimal.

The Flagstaff Limestone Formation is lithologically heterogeneous and possesses few recurrent features which might be said to be distinctive of it. Its separation as a formation rests not so much upon its own internal characteristics, but rather its contrast with those of underlying and overlying formations. Thus it differs from the Careg-onen Limestone Formation in the absence of porcellenous calcite mudstones, the lateral continuity of individual beds, the extensive development of rubbley limestones and the ready identification of palaeokarstic surfaces allowing its subdivision into minor cycles. Its varied suite of lithologies also contrasts with the lithologically monotonous cycles which comprise the overlying Moelfre Limestone Formation.

The Flagstaff Limestone Formation in Penmon is subdivided into the following minor cycles:-

Number	Average Thickness (m)
F6	4.00
F5	7.00
F4*	15.50
F3	2.00
F2*	8.00
Fl	0.50
TOTAL	37.00

* = There is a possibility that these minor cycles could be further divided and therefore represent multiple cycles. No definite palaeokarstic features to allow such subdivision have been recognised however.

The formation marks the first major development in the Anglesey succession of the enigmatic rubbly limestones. In the Flagstaff Limestone Formation such beds are rich in chonetid brachiopods.

In the lower cycles (Fl to F3) these rubbley limestones alternate with more massive to cross bedded sketetal grainstones. This alternation takes place on several scales. In general, the thinly bedded rubbley limestones are confined to and tend to predominate in the lower portions of these cycles. More massive to cross bedded skeletal grainstone units form the upper parts on which the palaeokarstic surfaces are developed. Thus minor cyclicity is demonstrated not only by palaeokarstic features but also by the distribution of constituent lithofacies. Following on from this, distinction between the two lithofacies on a smaller scale suggests a composite nature for cycle F2, but the absence of palaeokarstic features precludes its further subdivision (see Section 5.6). Rubbly units do occur in higher cycles in the same relative positions, but tend to be thin and laterally discontinuous. A notable exception is cycle F5 in which the lower 2 m are of this lithofacies.

Cycle F4 is conspicuous for the development of a thick (over 2.50 m) mudstone unit which forms an important marker horizon within the formation. The underlying cross bedded skeletal grainstones are highly fossiliferous and, towards the top, colonies of <u>Lithostrotion martini</u> (Edwards and Haine) are often so densely packed as to form a traceable coral band. The nature of the contact between this coraliferous grainstone and the mudstone unit remains debatable. The upper surface of the grainstone is gently undulating and is overlain by 10 to 12 cms of pale grey mudstone very similar to the bentonitic clays which often overlie palaeokarstic surfaces. There is, however, no suggestion of a laminated micritic crust or rootlets. By comparison with the Principal Area (see below) both units are regarded as belonging to the same cycle. The cycle is capped by a thick unit of cross-bedded, skeletal grainstones, oolitic in parts and with lenses of oncolitic coquina often rich in <u>Koninckopora</u> debris.

Cycle F5 as already mentioned, returns to the style of the lower cycles in the formation with its development of rubbly limestones.

The topmost cycle of the Flagstaff Formation, F6, is only well exposed in the cliffs near Fedw-fawr. Lower, thinly bedded and rubbley and brecciated limestones are overlain by an upper highly fossiliferous grainstone unit containing, in particular, many Palaeosmilia murchisoni (Edwards and Haime).

The palaeokarstic surface at the top of cycle F6 also marks the top of the Flagstaff Limestone Formation and is a particularly noteworthy one. Sandstone filled channels incised through this surface are up to 23 m deep impinging on the thick mudstone bed in cycle F4. Both the Fedw and Parc Sandstones represent such channel-fill clastics at the top of the Flagstaff Limestone Formation in the Penmon Area.

(ii) Principal Area (Fig.11; Chart 3)

In the Principal Area the Flagstaff Limestone Formation embraces the minor cycles which form the cliffs between Lligwy Bay and Porth Helaeth and comprises approximately 38 m of strata.

Inland exposure of the formation is poor, but it is clear from mapped relationships that, laterally, the lower cycles pass rapidly into the basal Lligwy Sandstone Formation. Upper cycles have been traced beyond Bwlch-y-dafarn [SH 4905 8565] south-west of which exposure fails but they too must be progressively lost to the diachronous sandstone facies. Certainly near Ponciau [SH 4760 8320] the Flagstaff Limestone Formation has pinched out completely as units of the succeeding Moelfre Limestone Formation in turn give way to the basal clastics (Fig.8).

The lithologically heterogeneous nature of the formation as in Penmon, contrasts markedly with the lithologically monotonous cycles of the overlying Moelfre Limestone Formation. Palaeokarstic

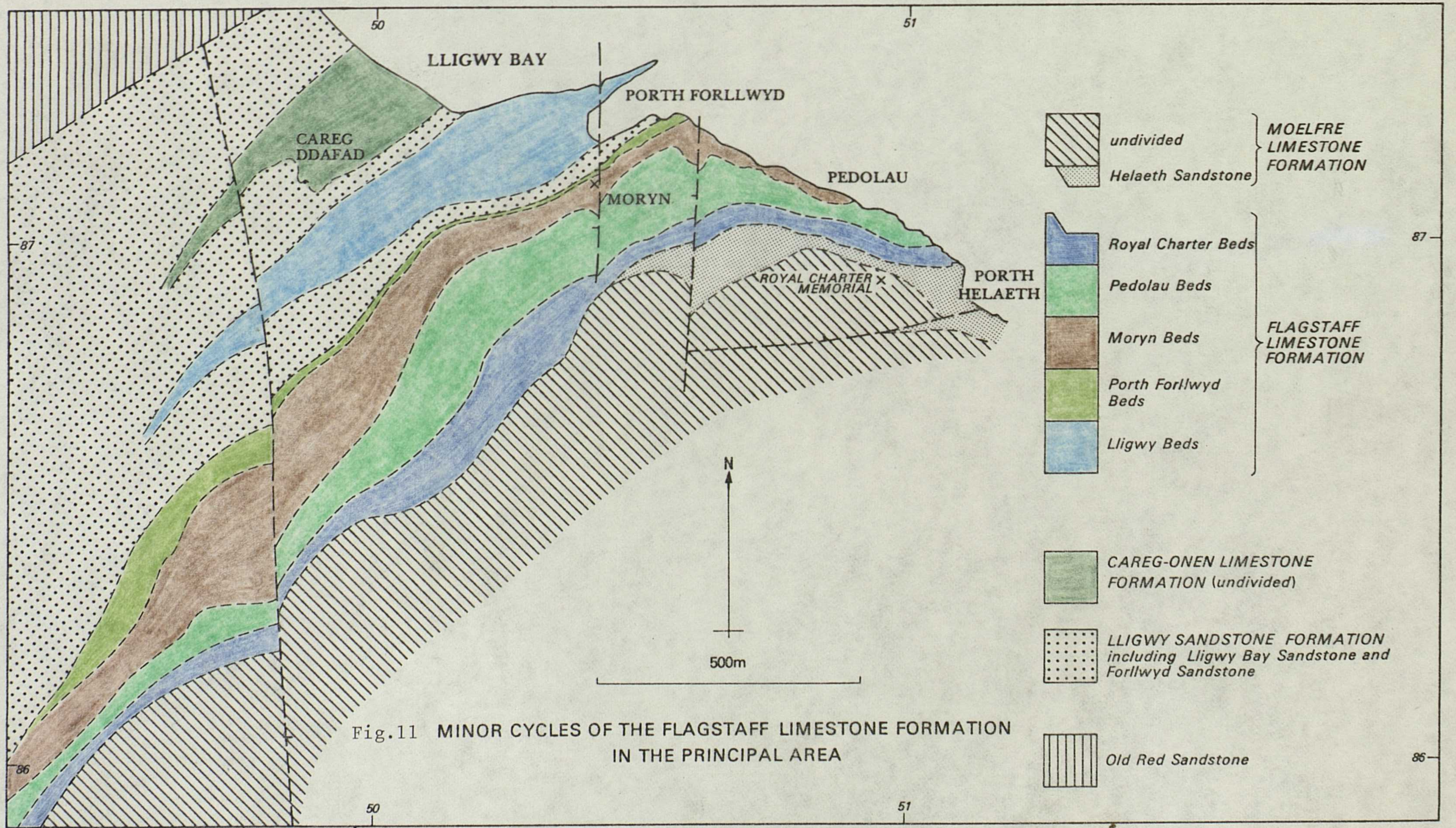


Fig.11 MINOR CYCLES OF THE FLAGSTAFF LIMESTONE FORMATION IN THE PRINCIPAL AREA

surfaces allow its ready subdivision into minor cycles. The extended coastal crop further allows the naming of individual cycles after type sections, rather than numbering (the benefits of such practice have already been discussed, Section 2.4). The Flagstaff Limestone Formation in the Principal Area is therefore subdivided into the following minor cycles:-

Name	Average Thickness (m)
Royal Charter Beds	5.00
Pedolau Beds	7.00
Moryn Beds	18.00
Porth Forllwyd Beds	3.00
Lligwy Beds*	5.00
TOTAL	38.00

* = suggested multiple cycle.

The occurrence of marker horizons within the formation allow the ready correlation of minor cycles between the Penmon and Principal Areas (Fig. 7). The thick mudstone unit in Penmon cycle F4 equates with that in the Lower Moryn Beds and is the single most important marker bed in the formation. The thick sequence of cross bedded grainstones which cap F4 in Penmon correlate with the similarly grainstone dominated Moryn Beds. The thinly bedded rubbley limestones which comprise the lower portion of F5 correspond to similar units in the Pedolau Beds.

Correlation below the Lower Moryn Beds is clearly more difficult with 3 minor cycles in Penmon and only 2 in the Principal Area. These problems are likely to be accounted for in the probable composite character of the Lligwy Beds. Differences in facies between the two areas in these lower cycles, including the lower part of the Moryn Beds are interpreted as being due to the proximity of the clastic shoreline as represented by the top of the Lligwy Sandstone Formation. This is borne out by the occurrence of the Forllywd Sandstone and a further thin sandstone towards the top of the Lligwy Beds. No clastics have been recorded from the equivalent cycles in Penmon.

The topmost cycle represented by the Royal Charter Beds correlates with cycle F6 in Penmon and is similarly fossiliferous, again with abundant P.murchisoni. On the coastal crop the top of the cycle and therefore the top of the formation is marked by the channel-filling Helaeth Sandstone. This therefore corresponds to the Fedw and Parc Sandstones of Penmon and further suggests that deposition of the Flagstaff Limestone Formation was terminated by a particularly marked regressive event, possibly of widespread correlatable value.

(iii) Age of the Flagstaff Limestone Formation

The formation contains a typical Asbian fauna including Palaeosmilia murchisoni (Edwards and Haine), Lithostrotion martini (Edwards and Haine), Dibunophyllum bourtonense (Garwood and Goodyear) and Koninckopora sp. Greenly took the top of the formation in the Principal Area as the top of D_1 subzone i.e. top of Asbian, but in fact the overlying Moelfre Limestone Formation is also regarded here as Asbian in age. The Flagstaff Limestone Formation is therefore of middle Asbian age.

(d) Moelfre Limestone Formation

The Moelfre Limestone Formation succeeds the Flagstaff Limestone

Formation in both the Principal and Penmon Areas and is the lowest limestone formation identified with confidence in the Straitside Area.

(i) Principal Area (Fig.12; Chart 4)

The formation is best seen overlying the Helaeth Sandstone in the north-facing cliffs of Porth Helaeth, north of Moelfre and this provides the type section. The uppermost parts of the formation and the contact with the overlying Traeth Bychan Limestone Formation are well seen on either side of Moelfre harbour. Faulting prevents an accurate estimate of thickness for the formation, but a minimum value is 32.5 m and it is unlikely to exceed 35 m. The formation is easily traced inland, but largely on features with few sections of note other than those provided by quarries overlooking Ponciau [SH 4810 8410]. As noted above, near here the Moelfre Limestone Formation overlaps the Flagstaff Limestone Formation and begins to pass into the Lligwy Sandstone Formation. The uppermost units of the Moelfre Limestone Formation and the contact with the overlying Traeth Bychan Limestone Formation have been identified in a series of small quarries in the ridge extending south from Yr-efail [SH 4810 8095] to Bryn-llwyd [SH 4795 7950]. These form the final exposures in the formation south of which it must soon be overlapped (Fig.8).

Further coastal exposures in the Moelfre Limestone Formation occur north of Benllech where it is brought in on the up-thrown side of the Huslan Fault (Fig.13). Up to 15 m of strata are attributable to the formation. These exposures are confined to the coastal strip with the overlying Treath Bychan Limestone Formation appearing immediately inland. The contact between the two formations is seen in a small cliff section in the car park at [SH 5218 8268].

In its type locality the Moelfre Limestone Formation is

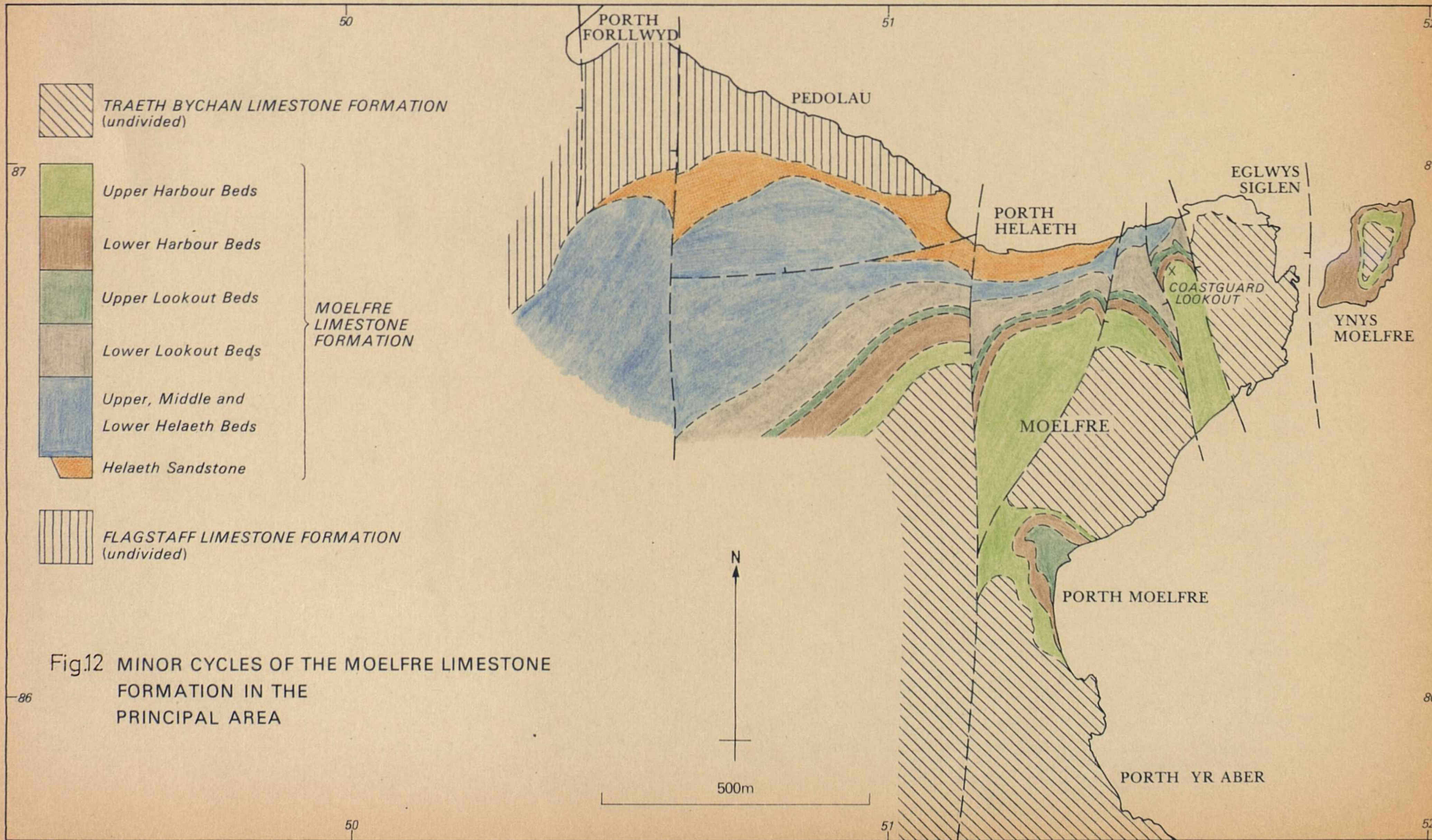


Fig.12 MINOR CYCLES OF THE MOELFRE LIMESTONE FORMATION IN THE PRINCIPAL AREA

distinctive in the lithological monotony of its constituent minor cycles contrasting with the more varied style of the underlying and overlying formations. The minor cycles are constructed virtually exclusively from massive, pale grey, skeletal grainstone/packstones (To classify rocks as lithologically monotonous clearly depends on the level of differentiation. Thus these cycles are monotonous in terms of the broad groupings of lithofacies, see Chapter 6, as recognised in the field). It is this lithological homogeneity which enables the formation to form bold inland escarpments and these in turn allow its ready mapping inland.

Despite their uniform lithological aspect minor cycles are readily identified in the formation. The varied palaeokarstic phenomena (hummocky surfaces, sandstone filled pots, laminated micritic crusts, brecciation etc.) contrast with the massive limestones and have allowed the following cycles to be recognised:-

Name	Av. Thickness (m)
Upper Harbour Beds	2.5
Lower Harbour Beds	3.0
Upper Lookout Beds	3.0
Lower Lookout Beds*	9.0
Upper Helaeth Beds	5.0
Middle Helaeth Beds	3.0
Lower Helaeth Beds*	7.0
TOTAL	32.5

* = possible multiple cycles.

Palaeokarstic surfaces within the formation have been traced laterally and observed to pass into stylolites with the loss of definitive palaeokarstic features. This poses two opposing questions: How many stylolites may be pressure dissolved palaeokarstic surfaces? Are all palaeokarstic surfaces truly of widespread lateral extent? These questions are discussed more fully in Chapter 3, Section 2 , but they suggest that either some of the above minor cycles may be composite in character, or alternatively that some could perhaps be grouped together.

Any individual description of the minor cycles at this stage is rendered redundant by their almost identical style. Of note however is the occurrence at the base of the Lower Lookout Beds of a thin (up to 50 cms) calcite mudstone bed with birdseye structures which pinches out against the swells in the underlying palaeokarstic surface. The occurrence of this lithology in these coastal exposures assumes greater significance when the inland sections are considered below.

As the escarpment which the formation forms is traced inland to the southwest the first sections of note are at Bwlch-y-dafarn [SH 492 855]. Here an area of conglomerate and sandstone has been mapped (also recognised by Greenly and by Mitchell) and although contacts with the surrounding limestones are not well seen it is likely that it represents a channel fill. Its top lies approximately 17 m below the top of the formation and therefore equates with the palaeokarstic surface at the top of the Upper Helaeth Beds. This has sandstone filled pots at the coast. Above the sandstone are limestones rich in Striatifera sp. and in the slopes above are beds of calcite mudstone with birdseyes. These are overlain by approximately 5.00 m of the standard skeletal grainstone/packstones extending to the top of the formation exposed in the small quarry at [SH 4958 8518] near Glan'r-afon. In the quarries overlooking Ponciau a thick (over 3 m base not seen) run of calcite mudstones with birdseye structures and rootlets occurs about 7.00 m below the top of the formation and would appear to be laterally equivalent to the Lower Lookout Beds at the coast. A higher calcite mudstone unit occurs 5 m below the top of the formation and equates with that observed at Bwlch-y-dafarn, horizons which equate to the Upper Lookout Beds. It is clear that a lateral facies variation is taking place within the formation. Certainly within the Lower and Upper Lookout Beds there is a transition from dominantly skeletal grainstone/ packstone cycles at the coast to ones in which calite mudstones constitute an important element. Such a variation is reminiscent of that in the Careg-onen Limestone Formation in Penmon. Significantly in the Moelfre Limestone Formation this variation is seen to take place as the clastic shoreline, as represented by the Lligwy Sandstone Formation, is approached.

(ii) Penmon Area (Fig.10; Chart 5)

The Moelfre Limestone Formation in the Penmon Area has been recognised in the steep cliffs east of Fedw-fawr and also in the upper of the Penmon Park Quarries. The degree of faulting in the area and the lack of any distinctive marker horizons in the formation has precluded a detailed appraisal of its distribution inland. For these same reasons an accurate assessment of the thickness of the formation is difficult. The thickest run of unbroken strata is 28 m and this embraces six minor cycles and the lower part of a seventh. With a further 2.75 m thick cycle present beneath the base of the overlying formation the thickness of the Moelfre Limestone Formation appears closely comparable to that in the Principal Area. However the recognition of palaeokarstic features has allowed further subdivision of the formation in Penmon than was allowed in the Principal Area and one, possibly two, further cycles may be identified.

Again the formation is characterised by massive skeletal grainstone/packstones from which the minor cycles are almost exclusively formed. Again the palaeokarstic surfaces are conspicuous, the more so in Penmon, because many are overlain by quite thick and often vividly varigated palaeosols. These allow the formation to be subdivided into the following minor cycles:

Number	Average Thickness (m)
M8	2.75
M7	2.00 (estimated)
M6	2.00
M5	4.75
M4	4.50
M3	3.50
M2	4.00
M1*	8.25
TOTAL	31.75

* = possible multiple cycle

A suggested correlation between these minor cycles and those of the Principal Area is given in Fig.7. Clearly in the absence of distinctive marker horizons this is based solely on thickness considerations and must therefore be treated with caution. It appears that minor cycles M4 and M5 equate with the Lower Lookout Beds for which a composite character was suspected in the Principal Area.

Again the uniform character of the minor cycles precludes their detailed description at this stage. No lateral facies variation as noted in the Principal Area has yet been recognised for the Moelfre Limestone Formation in Penmon.

(iii) Straitside Area (Fig.9)

Approximately 30 m of massive skeletal grainstone/packstones which outcrop below recognisable Brigantian strata on the Arfon side of the Straitside Area may, with some confidence, be equated with the Moelfre Limestone Formation. Poor exposure prevents their subdivision into minor cycles.

A lateral facies variation, similar to that seen in the Principal Area, does appear to take place in the Straitside Area also. As the massive limestones beneath the base of the Brigantian strata are traced westwards towards the Bridges Sandstone Formation units of calcite mudstone with birdseyes appear.

(iv) Age of the Moelfre Limestone Formation

Diagnostic macrofauna is absent from the formation and it was these strata which Greenly (1919 p.631) understandably regarded as transitional between D_1 and D_2. In thin sections of units from the formation fragments of Koninckopora sp have been identified and this calcareous algae is confined to rocks of Asbian or older age (Strank,1981). Foraminifera identified from the formation are also of Asbian affinity (Strank pers. comm.). Both Nichols and Mitchell assigned these strata to their foram zone 7, which may now be equated with the Asbian.

Overlying strata contain both macro- and microfauna which are diagnostic of the Brigantian stage. The Moelfre Limestone Formation is accordingly regarded as late Asbian in age and its top, therefore, markes the local position of the Asbian/Brigantian stage boundary.

(e) Traeth Bychan Limestone Formation

The Traeth Bychan Limestone Formation succeeds the Moelfre Limestone Formation in all three areas. It is the highest of the formations present in the Penmon Area.

(i) Principal Area (Fig.13; Chart 6)

The type section for the formation is the virtually continuous and readily accessible cliff line extending from Moelfre Harbour to the fault at Huslan, centred on Traeth Bychan. This comprises 96.5 m of strata. The formation succeeds the Moelfre Limestone Formation south of the Huslan Fault and forms much of the Thelwal headland to the south of Benllech. The formation outcrops extensively inland south of the main escarpment of the Moelfre Limestone Formation, and despite severe faulting individual cycles can often be identified and mapped. Southwest of Llanddyfran, however, exposure begins to fail, tracing of cycles is precluded and the formation as a whole is mapped with difficulty. It is in this ground, unfortunately, that the formation oversteps the underlying Moelfre Limestone Formation and starts to pass into the basal sandstone facies. The Pencraig Sandstone exposed about Llangefni is thought to correlate with the Benllech Sandstone (see below) at the coast. If so the highest units of the formation persist beyond Llangefni and must finally pinch out around Llan-fawr. Here the basal sandstone formation, of which the Pencraig Sandstone forms the local representative, has itself been overlapped and the limestones rest unconformably directly on the Mona Complex.

The Traeth Bychan Limestone Formation contrasts with both

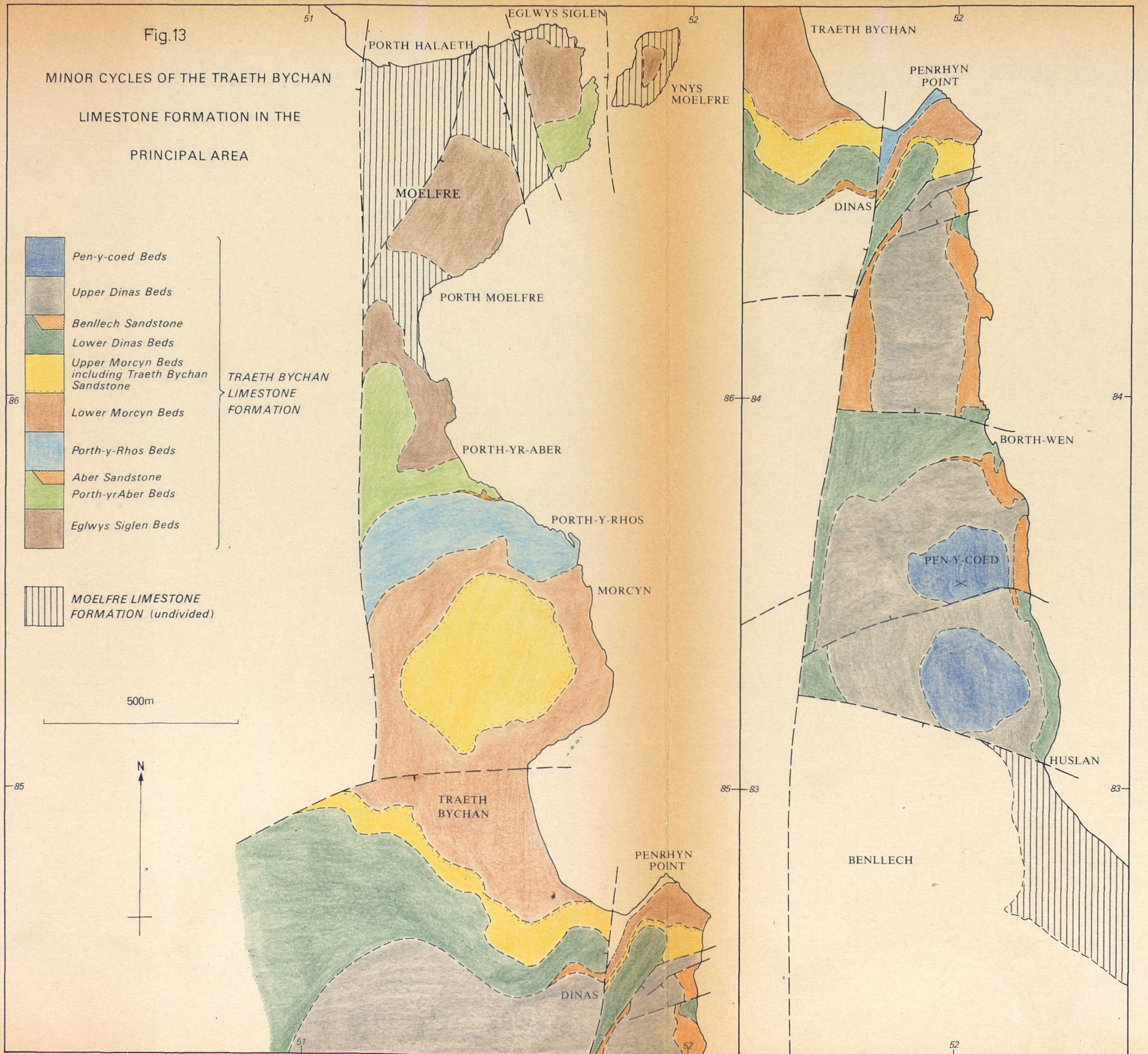


Fig.13
MINOR CYCLES OF THE TRAETH BYCHAN
LIMESTONE FORMATION IN THE
PRINCIPAL AREA
Pen-y-coed Beds
Upper Dinas Beds
Benllech Sandstone
Lower Dinas Beds
Upper Morcyn Beds including Traeth Bychan Sandstone
Lower Morcyn Beds
Porth-y-Rhos Beds
Aber Sandstone
Porth-yrAber Beds
Eglwys Siglen Beds
TRAETH BYCHAN LIMESTONE FORMATION
MOELFRE LIMESTONE FORMATION (undivided)
500m
N
51
52
86
85
84
83
EGLWYS SIGLEN
PORTH HALAETH
YNYS MOELFRE
MOELFRE
PORTH MOELFRE
PORTH-YR-ABER
PORTH-Y-RHOS
MORCYN
TRAETH BYCHAN
PENRHYN POINT
DINAS
BORTH-WEN
PEN-Y-COED
HUSLAN
BENLLECH

underlying and overlying formations in displaying a style of minor cyclicity emphasized not only by the occurrence of palaeokarstic surfaces, but by the ordered and predictable arrangement of the constituent lithofacies. The formation is also distinctive in that, amongst these lithofacies, dark argillaceous wackestones and packstones with associated shales form an important element. This argillaceous phase manifests itself in several ways. The thicker cycles tend to display thick runs of dark, thinly bedded, often tabular, wackestones and packstones with intercalated shale bands 2 to 3 cms thick. In others the phase is represented by a dark grey shale with limestone nodules many of which are clearly early diagenetic growths around burrows. Gradation between, and inter-digitation of these end members is common. Not all the minor cycles in the formation in the Principal Area develop this argillaceous phase. In those that do it always forms the lower portion of the cycle and gives way upwards to capping units of pale, more massive skeletal packstones and grainstones occasionally oolitic and cross bedded.

Chert, in the form of irregular nodules, becomes increasingly important towards the upper parts of the formation, but is largely restricted to the argillaceous limestones.

The formation has been subdivided into the following minor cycles:

Name	Average Thickness (m)
Pen-y-coed Beds	12.00
Upper Dinas Beds	13.50
Lower Dinas Beds*	19.00
Upper Morcyn Beds inc. Traeth Bychan Sst	3.00
Lower Morcyn Beds*	10.00
Porth y Rhos Beds	9.50
Porth yr Aber Beds	17.00
Eglwys Siglen Beds*	8.50
TOTAL	92.50

* = possible multiple cycles

Of these the Porth yr Aber, Lower Morcyn and the Lower and Upper Dinas Beds and the Pen-y-coed Beds exhibit the distinctive style mentioned above.

The lowest of the minor cycles, the Eglwys Siglen Beds, the base of which is also taken as the local base of the Brigantian, whilst not displaying this style nevertheless contrasts with the underlying units of the Moelfre Limestone Formation. It is lithologically more varied displaying an impersistent bed of oncolitic packstone and grainstone at its base (this has a thin development of laminated micritic crust at its top and should perhaps be regarded as a thin minor cycle in its own right), a conspicuous rubbley unit towards the middle of the cycle and capping beds of coarse crinoidal packstone. The cycle is distinctive in the abundance of Lithostrotion junceum (Flem.) and represents the "junceum-limestones" of Greenly (1919 p.632). These corals often occur in discrete bands of great value in local correlation. The occurrence of Nemistium edmondsi confirms a Brigantian age for these strata.

The Porth-yr-Aber Beds exhibit to perfection the lithofacies distribution so typical of the formation. Of note is the occurrence of beds rich in the large foraminiferan *Saccaminopsis* sp. within the thinly bedded argillaceous limestones. At Porth-yr-Aber the cycle is overlain by the Aber Sandstone which fills deep cylindrical potholes in the palaeokarstic surface.

The Porth y Rhos Beds are atypical of the formation, at least in the Principal Area, in returning to the style of the minor cycles of the underlying Moelfre Limestone Formation. However there is a little more lithological variety with the massive skeletal packstone/grainstones being supplemented by lower coquinoid limestones and capping ooid grainstones.

The Lower Morcyn Beds return to the characteristic style, the argillaceous phase is generally nodular or irregularly bedded rarely developing the discrete tabular beds of limestone typical of the thicker minor cycles. The Lower Morcyn Beds are particularly noteworthy for the Morcyn Coral Bed (Greenly, 1919 p.632 and Neaverson, 1946 p.135). This is not a bed at all in fact, but a zone towards the top of the argillaceous phase in which corals are variably concentrated, locally developing a boundstone texture. Common forms include *Lithostrotion irregulare*, *Diphyphyllum lateseptatum* and *L.maccoyanum*. The overlying massive skeletal packstone/grainstones represent the "*Lonsdaleia floriformis* Beds" of Greenly (p.632) in which the numerous colonies of that coral are often selectively dolomitised and form distinctive orange-brown patches on the rock surfaces.

The Upper Morcyn Beds have a distinctive basal sandstone bed up to 45 cms thick which often exhibits the trace fossil *Zoophycus*. This bed forms a valuable marker horizon in the coastal exposures both in the type section and in the Thelwal headland. Overlying thinly bedded rubbley limestones with clotted texture contain the brachiopod *Pleuropugnoides pleurodon* (Phillips) in abundance. The capping units of skeletal packstone/grainstone, like the underlying Lower Morcyn Beds, contain dolomitised corals with *Palaeosmilia regia* (Phillips) common.

The Lower Dinas Beds exhibit well the characteristic style of the thicker minor cycles of the formation. The palaeokarstic surface at the top of the cycle is of some importance since the Benllech Sandstone is confined in deep channels incised through this surface. Spectacular exposures of this sandstone are seen in the cliffs between Traeth Bychan and Benllech and it also outcrops extensively inland. Further description and interpretation of the often complex channel filling sequences are given in Chapter 5, Section 5.6b. Arguably, as with the Helaeth, Fedw and Parc Sandstones at the top of the Flagstaff Limestone Formation, the widespread development of the Benllech Sandstone might indicate a major regressive event. The overlying Upper Dinas Beds, however, form a minor cycle identical in style to both the Porth-yr-Aber and Lower Dinas Beds and, outside the channel complexes, its inclusion in the same formation would be undoubted. In such areas the palaeokarstic surface is not especially pronounced, although at Huslan it is of note for the occurrence of slot shaped borings assigned to the ichnogenus *Trypanites* in addition to the more normal palaeokarstic phenomena. Subsequent studies beyond the Anglesey crops may indeed establish this horizon as a widely correlatable emergent episode of mesothemic significance. For the time being however the lithostratigraphic approach adopted herein dictates that the Upper Dinas Beds and indeed the overlying Pen-y-coed Beds be included in the Traeth Bychan Limestone Formation.

The Upper Dinas Beds are of further note for the superb coral beds which they exhibit near Penrhyn Point [SH 5196 8455] composed of the intermeshed and silicified colonies of Lithostrotion and Diphyphyllum, figured by Greenly (1919, plate XXXVII). Individual beds die out quite rapidly when traced laterally, but the zone as a whole, within the lower thinly bedded argillaceous limestone phase, is always rich in corals (c.f. Morcyn Coral Bed).

The Pen-y-coed Beds form the two small outliers which cap the cliffs between Borth-wen and Huslan. This is the least well exposed of the minor cycles on the coastal crop, but clearly exhibits the typical style of the formation. It is distinctive in that the upper units of skeletal packstone and grainstone are locally sandy with stringers of pebbley sandstone.

The similarity of style of the upper three minor cycles often makes their distinction difficult in inland areas where exposure is poor and/or faulting is pronounced. This is particularly marked along the southern margin of the Principal Area where there are thick drift deposits and faulting becomes increasingly intense as the Berw Fault complex is approached. Unfortunately it is in this ground that the contact between the Traeth Bychan Limestone Formation and the succeeding Red Wharf Cherty Limestone Formation must be defined since it is not exposed at the coast. Exposures around Betws [SH 510 813] suggest that the Red Wharf Cherty Limestones rest on the Pen-y-coed Beds. It may be however that there are further higher minor cycles unexposed or as yet unrecognised which belong to the Traeth Bychan Limestone Formation. For the moment this remains an area of uncertain stratigraphy in the succession.

Where traceable inland the cycles exhibit no significant changes in either facies or thickness. The critical parts of the cycles where they impinge upon the basal sandstones are, unfortunately, never well exposed.

(ii) Penmon Area (Fig.10; Chart 7)

The Traeth Bychan Limestone Formation has been recognised in numerous fault bounded blocks along the northern coast of the Penmon Area. The fullest sequences are observed at Nant Franen, the cliffs and quarries east of Fedw Fawr and towards the eastern point both at Trwyn-du and Trwyn Dinmor. In the latter locality large quarries immediately inland provide further sections through the formation. The degree of faulting in the area has, to date, prohibited a detailed assessment of its inland distribution.

The top of the uppermost minor cycle, equivalent to the Upper Dinas Beds of the Principal Area, has not been observed and so the true thickness of the formation in Penmon cannot be measured. A compilation thickness for the exposed part of the formation is 76.5 m and indicates a slight thickening compared with the Principal Area.

The formation in Penmon exhibits the characteristic argillaceous lithologies and style of cyclicity described from the Principal Area. Indeed every minor cycle conforms to the standard pattern in some degree. The following minor cycles have been identified:

Number	Thickness (m)
TB 7 (top not seen)	6.00 (observed)
TB 6*	16.50 (estimated)
TB 5	5.00
TB 4*	9.50
TB 3	7.50
TB 2	19.00
TB 1*	8.50
TOTAL	72.00

* = possible multiple cycles

These minor cycles are readily correlated with those observed in the Principal Area (Fig.7).

The lowest minor cycle, TB 1, is readily distinguished from the underlying Moelfre Limestone Formation. The cycle contains a 2 m thick bed of dark grey shale with abundant nodules after burrows apparently the lateral equivalent of the rubbley zone described from the Eglwgs Siglen Beds in the Principal Area. This is a variety of the argillaceous phase typical of the cycles in this Formation but TB 1 differs from the normal style in that this phase does not form the lowest part. It is separated from the nearest underlying palaeokarstic surface by 2 m of skeletal packstone/grainstone similar to the underlying units of the Moelfre Limestone Formation. The oncolitic units variably developed at the base of cycle in the Principal Area appear to be absent in Penmon, but on thickness considerations below the rubbley zone the defining palaeokarstic surfaces correlate quite closely. The cycle is rich in Lithostrotion junceum (Flem), again often forming discrete coral bands.

The equivalent minor cycle to the Porth-yr-Aber Beds, TB 2, appears to be thicker in the Penmon Area, but again exhibits the same characteristic distribution of lithologies. It is in this minor cycle that the foraminiferan Saccaminopsis sp. makes its first appearance in the Penmon succession. The cycle forms the spectacular cliffs at Trwyn Dinmor figured by Greenly (1919, plate XLI).

In the Penmon Area the minor cycle equivalent to the Porth-y-Rhos Beds, TB 3, has a lower argillaceous phase of nodular shales which pass upwards into coarse coquinoid grainstones.

Minor cycle TB 4 resembles closely its equivalent in the Principal Area, the Lower Morcyn Beds, although the coral-rich zone towards the top of the argillaceous phase is not developed. The overlying skeletal packstone/grainstones are however similarly rich in dolomitised corals including Lonsdaleia floriformis (Flem).

The succeeding minor cycle TB 5 has only been observed in the cliffs and coastal quarries east of Fedw-Fawr and here it is not well exposed. Dark grey shales are developed towards the base and indicate the occurrence of a lower argillaceous phase differing from the contemporary Upper Morcyn Beds in the Principal Area. The shales which contain Pleuropugnoides plurodon (Flem) are overlain by pale, massive, skeletal packstones and the cycle therefore now conforms to the normal style.

Minor cycle TB 6 is similarly poorly exposed and has nowhere been observed in full. The lower shales with thinly bedded, dark, argillaceous wackestones overlie the palaeokarstic surface at the top of TB 5 in the section east of Fedw-Fawr. Upper units of the cycle are exposed beneath the succeeding minor cycle at Trwyn Du and consist of irregularly bedded argillaceous packstones with abundant <u>Lonsdaleia duplicata</u>, overlain by capping beds of skeletal grainstone. These latter units are oncolitic towards their base, often cross bedded and sandy in places, locally forming a calcareous sandstone. The palaeokarstic surface at the top of the cycle is equivalent to the Benllech Sandstone level in the Principal Area. It has a sandstone veneer, intraclast breccia and displays the borings <u>Trypanites</u> as well as more normal palaeokarstic phenomena. Thicker units of sandstone and conglomerate exposed immediately to the east may represent thickening of the sandstone veneer into a channel-fill sequence, but faulting obscures the exact relationships.

Overlying this palaeokarstic surface are 6 m of shales and dark, thinly bedded, argillaceous wackestones. These represent the only portion of minor cycle TB 7 exposed and are thought to be the highest Dinantian deposits in the Penmon Area.

(iii) <u>Straitside Area</u> (Fig.9; Chart 8)

Minor cycles attributable to the Traeth Bychan Limestone Formation in the Straitside Area have only been examined on the Anglesey side. They are exposed along the cliffs between Plas Newydd and Llwyn Chwarel-gôch [SH 5219 6860]. The following minor cycles, which display well the characteristic argillaceous phases have been identified:-

Name	Average Thickness (m)
Spring Beds	7.00
Upper Carnedd Beds inc. Carnedd Sst	13.00
Lower Carnedd Beds	11.00
Upper Dyke Beds	11.50
Lower Dyke Beds	2.00 (exposed)

The contact with the underlying Moelfre Limestone Formation is not exposed and correlation of these minor cycles with those of the Principal and Penmon Areas has not yet been achieved.

Only the upper portions of the Lower Dyke Beds are exposed. Shales with thin beds of coquinoid grainstone rich in Pleuropugnoides pleuòdon are overlain by coraliferous wackestones and the cycle is capped by a thin, rootlet infested, sandy limestone. The overlying mudstones of the Upper Dyke Beds also contain abundant Pleuopugnoides pleurodon and correspond to the "Pleurodon Mudstones" of Arvon described by Greenly (1928). A 1.50 m thick massive bed of skeletal packstone tops the cycle.

The Lower Carnedd Beds display the first of the two "great coral beds" described by Greenly (1919 p.649). Not a bed, but a coral rich zone towards the top of the interbedded argillaceous wackestones and shales which characteristically form the lower portions of the cycle (c.f. Morcyn Coral Bed).

The Carnedd Sandstone separates the Lower from the Upper Carnedd Beds which remain sandy throughout. The overlying Spring Beds contain the second of Greenly's coral beds with a style identical to that of the lower. The well developed ooid grainstones which cap this minor cycle represent the often quoted Edwen Oolite of Greenly (Neaverson, 1945; George, 1974). The palaeokarst surface which defines the top of cycle is well seen in the old quarry at Llwyn Chwarel-gôch. Overlying massive limestones with sandstone intercalations (the Edwen and Moel-y-don Sandstones) and containing abundant Lonsdaleia duplicata are thought to belong to the higher Red Wharf Formation.

(iv) Age of the Traeth Bychan Limestone Formation

The formation contains a typically Brigantian fauna including the diagnostic forms Lonsdaleia floriformis, Lonsdaleia duplicata, Dibunophyllum bipartitum, Palaeosmilia regia, Lithostrotion irregulare Diphyphyllym lateseptatum, Saccaminopsis sp., Caninia juddi.

The occurrence of Saccaminopsis sp. from the Porth-yr-Aber Beds is of particular interest from the point of view of regional correlation and is discussed more fully in Section 2.5.

(f) Red Wharf Cherty Limestone Formation

This formation is only described from the Principal Area. It is almost certainly absent in the Penmon Area. Massive and cherty limestones almost certainly represent units of the formation in the Straitside Area,(Fig.9, but to date these have not been examined in detail.

(i) Principal Area (Fig.14; Chart 9)

The cliff and quarry sections of the headland to the north of Red Wharf Bay form the type section through the formation. Faulting isolates these outcrops from the remainder of the Principal Area and since the base of the formation is not exposed its relationship to the underlying Traeth Bychan Limestone Formation is unclear. Inland mapping suggests that the lower units of the Red Wharf Cherty Limestone Formation rest on the Pen-y-coed Beds, but faulting and poor exposure make the correlation necessarily uncertain.

The minor cycles which comprise the formation are predominantly constructed from pale, massive, skeletal packstones and grainstones and contrast with the underlying minor cycles of the Traeth Bychan Limestone Formation in the absence of argillaceous limestone phases (c.f. Moelfre Limestone Formation). As its name suggests chert occurs in abundance throughout the Red Wharf Cherty Limestone Formation as both nodules and thin beds culminating in a 6 m thick sequence of bedded cherts which form the highest units of the Dinantian succession in Anglesey. These highest Dinantian strata are overlain by rocks of Upper Carboniferous (Silesian) age (Greenly, 1919; George, 1961).

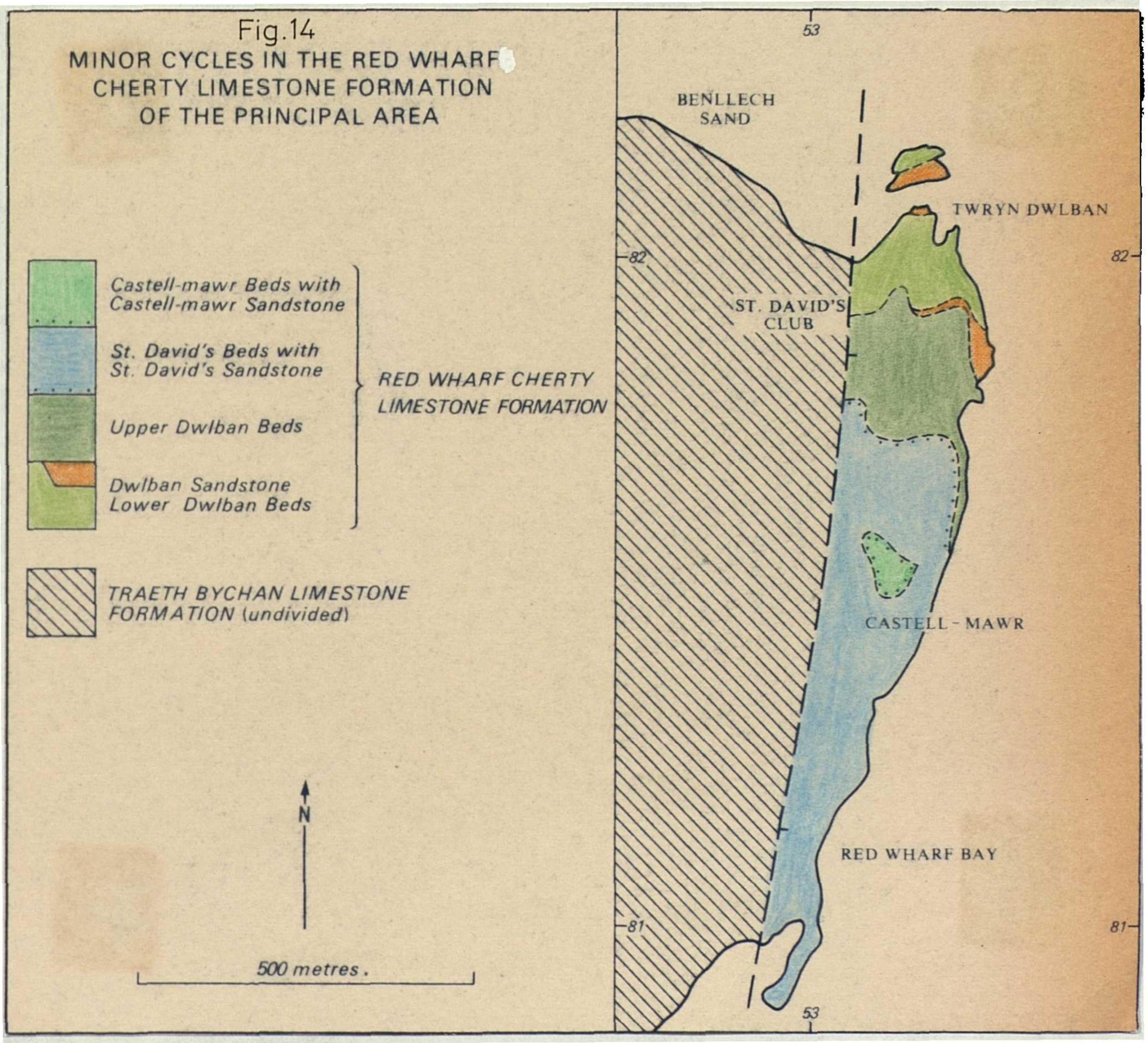


Fig.14
MINOR CYCLES IN THE RED WHARF
CHERTY LIMESTONE FORMATION
OF THE PRINCIPAL AREA

Often spectacular palaeokarstic surfaces allow the formation to be subdivided into the following minor cycles:-

Name	Thickness (m)
Castell Mawr Beds inc. Castell-mawr Sst	11.00
St. Davids Beds inc. St. Davids Sst	23.50
Upper Dwlban Beds	11.50
Lower Dwlban Beds	8.00 (exposed)
TOTAL	54.00

As mentioned above, the base of the Lower Dwlban Beds is not exposed and it may be that a lower argillaceous phase exists prompting this minor cycle's inclusion with the Traeth Bychan Limestone Formation. As it is the exposed parts of the cycle are wholly developed in irregularly bedded skeletal packstones, crinoidal towards the top with the ubiquitous chert nodules. The palaeokarstic surface at the top of this cycle exposed at Trwyn Dwlban displays the famous sandstone pipes (Section 3.2, Plates 12,13) frequently visited by student parties and referred to by several authors (Greenly 1900, 1919; Morton, 1901; Challenor and Bates, 1974 and Baughen and Walsh, 1980). A more detailed description and interpretation of the features this surface displays, noting in particular the several generations of fill, is presented by Walkden and Davies (in prep.). These siliciclastics which also fill channels in the surface represent the Dwlban Sandstone.

The Upper Dwlban Beds have irregularly bedded skeletal packstones in their lower part with abundant Lonsdaleia duplicata and Chaetetes sp. These give way to massive units of skeletal packstone/grainstone. The palaeokarstic surface at the top of this minor cycle is again remarkable displaying the largest sandstone filled pot-holes recorded from the succession (Plate 14) see also Greenly, 1919, plate XXXIX).

These fills are continuous with the overlying St. Davids Sandstone the basal member of the St. Davids Beds. This sandstone appears to have a sheet-like geometry as opposed to the more normal channel filling style. The remaining parts of this cycle are composed of a thick run of massive skeletal packstone/grainstones.

The Castell Mawr Beds which cap the conspicuous plateau-like feature of that name to the north of Red Wharf Bay have the pebbley and calcareous Castell Mawr Sandstone at the base. This gives way to massive skeletal packstone/grainstones which become increasingly cherty and pass into the topmost bedded cherts.

(ii) Age of the Red Wharf Cherty Limestone Formation

The occurrence of Lonsdaleia duplicata confirms a Brigantian age for these strata.

2.5 REGIONAL CORRELATION

The minor cyclic nature of Asbian and Brigantian strata throughout the country is now widely established (e.g. Ramsbottom, 1973; Walkden, 1974; Somerville, 1979a and c; Burgess and Mitchell, 1976). The correlation of individual minor cycles between these various sequences is a worthy goal but perhaps beyond the present state of knowledge with numbers holding sway over detailed lithological and palaeontological comparison (George, 1977). Correlation on a stage and formational basis may be attempted however and may in turn provide a framework for the future assessment of minor cycle equivalence.

Fig.15 illustrates probable lithostratigraphic correlations for the Anglesey sequence with Dinantian successions on the North Wales mainland. The formations recognised on Anglesey broadly equate with those erected by Somerville (1979a, b and c) in the Llangollen and

Regional Stages	ANGLESEY (this thesis)	GREAT ORME Morton (1898)	MOLD Somerville (1979c)	LLANGOLLEN Somerville (1979a,b)	LLANGOLLEN Morton (1870,1878)	PRESTATYN AND GLODDAETH SYNCLINE Nutt and Smith (in press)
BRIGANTIAN	RED WHARF CHERTY LIMESTONE FORMATION		Variable but including : Sandy Passage Beds, Arenaceous Limestones, Cefn-y-fedw Sandstone, Bedded Chert and Cherty Shale.			GRONANT GROUP
BRIGANTIAN	TRAETH BYCHAN LIMESTONE FORMATION	UPPER GREY LIMESTONE	CEFN MAWR LIMESTONE FORMATION	TREFOR LIMESTONE FORMATION	UPPER GREY LIMESTONE	
ASBIAN	MOELFRE LIMESTONE FORMATION	Upper Beds ? —— Lower Beds; MIDDLE WHITE LIMESTONE	LOGGERHEADS LIMESTONE FORMATION	EGLWYSEG LIMESTONE FORMATION	MIDDLE WHITE LIMESTONE	'knoll reef' limestone
ASBIAN	FLAGSTAFF LIMESTONE FORMATION					DYSERTH LIMESTONE GROUP
ASBIAN	CAREG-ONEN LIMESTONE FORMATION	LOWER BROWN LIMESTONE	LEETE LIMESTONE FORMATION	TY-NANT LIMESTONE FORMATION	LOWER BROWN LIMESTONE	'reef'
ASBIAN	BASAL SANDSTONES		BASEMENT BEDS	BASEMENT BEDS	BASEMENT BEDS	BASEMENT BEDS

Fig.15 Regional Correlation

Mold districts. Lower strata in these latter areas, the Leete and Ty-nant Formations, are rich in porcellaneous calcite mudstones with birdseye structures and correspond to the Careg-onen Limestone Formation in the Penmon Area. Somerville's Eglwyseg and Loggerheads Formations comprise massive bioclastic and rubbly limestones reminiscent of the Flagstaff and Moelfre Limestone Formations. Overlying Brigantian minor cycles which comprise Somerville's Cefn Mawr and Trefor Formations exhibit pronounced lithological rhythmisity and contain lower thinly bedded argillaceous limestone phases. These strata compare closely with minor cycles of the Traeth Bychan Limestone Formation and suggest that detailed cycle to cycle correlation may ultimately be possible. The highest Dinantian strata around Llangollen are sandy but pass laterally, in the Mold district, into chert rich units which broadly equate with the Red Wharf Cherty Limestone Formation.

Somerville's more up to date lithological descriptions of the sections in the Llangollen area also allow the more precise comparison of the Anglesey formations with the earlier colour based scheme of Morton since this was initially erected in the former district (Morton, 1870, 1878). Thus although Morton (1901) had difficulty in recognising his own divisions on Anglesey (Section 2.2) the formational groupings of minor cycles outlined above (Section 2.4) in fact reflect his three fold divisions of Lower Brown, Middle White and Upper Grey Limestones (Fig.15). This is significant since Morton's work remains to date the only attempt at regional lithostratigraphic correlation in North Wales. He was able to recognise his three fold divisions in many of the mainland sequences (Morton, 1886, 1897, 1898; see also Wedd, Smith and Wills, 1927) and this now suggests that the broad formational groupings established on Anglesey are of widespread significance.

Recent work by the Institute of Geological Sciences in the Denbigh and Rhyll areas (Nutt and Smith, memoir in press) has confirmed Morton's divisions within many of mainland sections which intervene between Anglesey and Llangollen. Significant variations have, however, been noted in the most northern crops around Prestatyn and the Gloddaeth Syncline (see also Neaverson, 1930, 1937 and George, 1974). Here Morton's three fold subdivision is not readily applied; there is pronounced increase in thickness and these strata are not obviously cyclic in character (author's field observation). Furthermore it is in these areas that knoll reef limestones occur (Neaverson, 1929), remnants of a linear belt of reefs which once defined the edge of the Dinantian shelf in North Wales (Fig.4, Section 1.4; Ramsbottom, 1969). The more complex sequences encountered in these latter areas appear, therefore, to be a reflection of their shelf edge and shelf slope settings, whilst widely correlatable formations and possibly minor cycles appear confined, palaeogeographically, to the shelf lagoon.

The lithostratigraphic patterns recognised in North Wales may also be compared with more distant Dinantian sequences (see Somerville, 1979a, b and c for more detailed discussion). Low Asbian strata which similarly display a considerable development of calcite mudstones have been recognised in Northern England (Ramsbottom, 1974). Higher Asbian strata comprising thickly bedded bioclastic limestones, but in which minor cyclicity is also evident have been described from the Malham district and from Derbyshire (Schwarzacher, 1958; Walkden, 1974). These appear closely comparable both in age and style with the minor cycles of the Moelfre Limestone Formation.

Brigantian strata in North Wales compare closely, perhaps significantly, with the widely correlatable Yoredale type cyclothems of the Alston and Askrigg Blocks (Ramsbottom, 1974; Burgess and Mitchell, 1976; Mitchell, 1978) and it is tempting to think that with perhaps more refined palaeontological investigation cycle to cycle correlation between the two successions may one day be achieved. It should always be borne in mind, however, that these Northern England sequences were laid down in a different tectonic province to the north of the intervening Craven Basin and that until the causative mechanisms for minor cyclicity are established detailed correlation with strata in North Wales, certainly in terms of event stratigraphy, may be unwarranted. In this context however it is interesting to note the widespread occurrence of units rich in the large foraminiferan *Saccamminopsis* in the Porth yr Aber Beds (Section 2.4(e)). Similar units have been recognised in the second minor cycle of Somerville's Cefn Mawr Formation in the Mold district (Somerville, 1979c) but also from equivalent low Brigantian Yoredale cyclothems in Northern England (Burgess and Mitchell, 1976) as well as in analogous strata in Derbyshire (George et al, 1976).

CHAPTER THREE

PALAEOKARSTIC SURFACES AND ASSOCIATED PHENOMENA

3.1 INTRODUCTION

It is clear from the preceding Chapter that the recognition, throughout the succession, of surfaces which record penecontemporaneous subaerial exposure has been of great importance. These surfaces form the basis for stratigraphic subdivision of the Dinantian sequence as well as providing invaluable environmental criteria. The surfaces record marine regressions, periods when the limestone strata were exposed and subject to lithification, solution and pedogenic alteration. Such surfaces are termed palaeokarstic (Walkden, 1974). Pedogenic and other emergent phenomena from both carbonate and siliciclastic deposits are the subject of an expanding field of research. Detailed description of the Anglesey palaeokarstic horizons has allowed their comparison with increasingly better documented modern equivalents and resulted in their more precise interpretation.

Previous workers on the Anglesey Dinantian have, in the main, failed to recognise these ancient emergent surfaces. The notable exception is Greenly's work on the sandstone pipe horizons within the succession (see Section 3.2d). These, he concluded, recorded periods ". . . of shallowing, and perhaps exposure of the sea-bottom" (Greenly, 1900 p.23). Greenly's studies of these particular phenomena have been extended by subsequent workers (Mitchell, 1964; Challinor and Bates, 1973; and Baughan and Walsh, 1980), but the vast majority of unpiped, more 'normal' palaeokarstic horizons have remained unnoticed. Palaeokarstic surfaces within other Dinantian sequences have been described from North Wales (Somerville, 1979a,b,c), from Derbyshire (Walkden, 1972, 1974, 1977; Adams, 1980), from South Wales (Riding and Wright, 1980; Wright, 1982) and from the U.S.A. (Walls et al, 1975; Harrison and Steinen, 1978).

Three features combine to make palaeokarstic horizons distinctive: 1) the morphology of the palaeokarstic surface; 2) associated overlying bentonitic palaeosols; and 3) the complex alteration phenomena developed in underlying host limestones.

3.2 MORPHOLOGY OF PALAEOKARSTIC SURFACES

(a) Description

The morphology of palaeokarstic surfaces can be expressed both on a regional scale and at outcrop level.

An assessment of the regional morphology of palaeokarstic surfaces comes from stratigraphic considerations and an appreciation of lateral thickness variation. Correlation of the various Dinantian sequences on Anglesey demonstrates relatively small thickness variations (Fig.6). Individual minor cycles in the Principal and Penmon areas exhibit only small changes in thickness rarely of more than 2 m, over distances of up to 15 km (Chapter 2). On a regional scale, therefore, palaeokarstic surfaces which define the minor cycles, appear to have been remarkably flat, even and laterally extensive.

At outcrop level palaeokarstic surfaces are characteristically highly uneven in form and consist typically of a series of bowl shaped or more rarely flat bottomed depressions or pits separated by irregular ridges and hummocks (Plate 3). The latter are characteristically scolloped in appearance and locally develop small overhangs. Average amplitudes are around 25-30 cms but surfaces exhibit a spectrum of morphologies ranging from gently undulating with amplitudes of less than 10 cms and locally flat, to surfaces where the depressions have developed into deep conical pits up to 2 m deep (Plate 4a,b). These latter features may in turn be related to the deep, cylindrical, sandstone filled pipes described below (Section 3.2d). Palaeokarstic surfaces are commonly smooth and markedly erosive truncating fossils in the underlying limestones. Contacts with overlying units are generally sharp and well defined. In other cases surfaces are more rubbly or brecciated and exhibit digitate contacts with succeeding beds. This range of morphologies is apparent not only throughout the succession, but may also be observed as individual surfaces are traced laterally. The typical hummocky surfaces and the more deeply pitted forms correspond to types 1 and 2 palaeokarstic surfaces of Walkden (1974) from the Dinantian of Derbyshire. His type 3 surfaces where the deeper pits coalesce forming isolated limestone pinnacles are not recognised on Anglesey.

Palaeokarstic surfaces are generally overlain by bentonitic palaeosols containing rootlets and occasional thin coals (Section 3.3) which thin or even pinch out against the higher hummocks and ridges. Where such palaeosols are thin or absent, overlying limestone or more rarely sandstone beds mantle the underlying irregularities (Plate 4c) and since such contacts are also boundaries of minor cycles they often coincide with pronounced changes in lithofacies.

The complex alteration phenomena, largely pedogenic in origin (Section 3.4), which characterise palaeokarstic horizons are generally only developed in limestones immediately (up to 2 m) below the surfaces, whilst lithoclasts of altered limestone occur in the basal beds of succeeding limestone cycles. Occasionally palaeokarstic surfaces give way laterally to stylolites and the characteristic hummocky form may be lost due to pressure solution. The truncated alteration textures then provide the only evidence of such a surface's former existence and of the position of the minor cycle boundary it once marked.

(b) Interpretation

The presence of alteration textures only immediately below palaeokarstic surfaces and the mantling nature of overlying limestones clearly demonstrate that the surfaces with their often pronounced hummocky topography were extant prior to the deposition of the succeeding strata.

Evidence of the early lithification of the limestones below palaeokarstic surfaces is provided by the steep and locally overhanging sides to the ridges and hummocks. Truncated fossils evidence the erosive origin of the surfaces, whilst the irregularity of the pits and the scolloped nature of the hummocks strongly suggests that this erosion was of a solutional rather than mechanical type (Walkden, 1974). Nor do the surfaces resemble the jagged and fretted morphologies of intertidal solution platforms ancient analogues of which have been described from Ordovician limestones by Read and Grover (1977).

The Anglesey surfaces more closely resemble karstic features developed in the Pleistocene limestones in the West Indies (Ruhe et al, 1961; Harrison, 1977). Piped, pitted and hummocky surfaces overlain by fossil soils are recorded from these limestones. These features were formed during Pleistocene low stands of sea level when the host limestones were subject to emergence, subaerial weathering and vadose diagenesis (Land et al, 1967). The pipes and pits formed as a result of dissolution of the limestones by rain water percolating downwards to lowered water tables. The soils include insoluble residues left after dissolution of the limestones (Blackburn and Taylor, 1969), but appear mainly to represent accumulations of wind blown atmospheric dust (Land et al, 1967; Harrison, 1977).

The role of a soil cover in karst formation has been discussed by Bronte (1963) and Sweeting (1972). Smooth irregular or rolling karstic surfaces result from solution beneath a soil cover (covered karst), whilst solution on uncovered limestone surfaces gives rise to delicately sculpted and jagged karst morphologies (uncovered karst). The Anglesey surfaces more closely resemble the former covered karst variety. The presence of overlying bentonitic palaeosols with rootlets and occasional coals supports such a model and confirms a palaeokarstic interpretation. The early lithification (cementation) of the limestones is therefore likely to have been related to subaerial exposure and an effect of vadose diagenesis. The petrographic evidence for early vadose cements is discussed in Section 3.4. Such lithification is analogous to the 'case hardening' suffered by Recent freshly exposed carbonate sediment (Bathurst, 1971, p.328). Following lithification dissolution at the base of overlying palaeosols by downward perculating rainwaters took place and the characteristic irregular, pitted and hummocky morphologies were generated.

In an extension of this model Wright (1982) has drawn attention to the similarities of hummocky (he uses the term 'mammillated' after Walkden, 1974) palaeokarstic surfaces in the British Dinantian to Recent Makondo Karst recognised in South Africa (Sweeting, 1972, p.94). Individual pits and hollows of this modern covered karst are related to trees rooted in the overlying soil. The pits form in response to localised biochemically controlled dissolution around each root mass.

Walkden (1974) has discussed rates of solution and length of emergence for Dinantian palaeokarst in Derbyshire. By comparison with limestone dissolution rates under analogous present day climatic regimes i.e. humid tropical and subtropical oceanic, he concluded that Derbyshire surfaces of average amplitude (i.e. 50 cms) could mark emergent episodes of between 30,000 and 100,000 years duration. The validity of such calculations is questionable especially since the average amplitudes differ not only between Anglesey and Derbyshire, but also from outcrop to outcrop on a single surface.

Whilst there are close similarities between the Anglesey palaeokarstic surfaces and Pleistocene covered karst phenomena at outcrop level there are serious discrepancies when the broader regional properties are compared. Bretz (1960) and Land et al (1967) have discussed the regional morphology of the Bermuda palaeokarsts; these exhibit pronounced relief of over 30 m. The packages of sediment which they bound are highly irregular pinching out rapidly as they are traced laterally and contrast with the lateral continuity of Dinantian minor cycles, both on Anglesey and elsewhere in North Wales (Somerville, 1979). Lateral impersistence of disconformity bounded units has also been noted from Recent and Pleistocene limestones on Aldabra (Braithwaite, 1975) and from Shark Bay (Read, 1974).

Pleistocene carbonate sequences record alternations of marine sedimentation and subaerial exposure, a response to eustatic changes in sea level related to glacial and interglacial periods. However the Pleistocene limestones described above, which are exposed today during the present high stand of sea level, represent atypical marginal sequences deposited at the extremities of Pleistocene transgressions. These sequences tend to be thin and attenuated with closely spaced discontinuity surfaces, whilst the often complex regional geometry of these surfaces reflects the repeated, often prolonged periods of emergence and karstification experienced in such settings.

Buried Pleistocene palaeokarstic surfaces described by Perkins (1977) from Florida appear more closely comparable to the Anglesey examples both in terms of regional morphology and gross setting. These are developed in mixed carbonate/siliciclastic lagoonal sequences which accumulated behind shelf edge coralgal reef and high energy shoal complexes. Boreholes through these Florida sequences have allowed the recognition and widespread correlation of five laterally persistent sedimentary units, Q1 to Q5, each defined by palaeokarstic 'discontinuity' surfaces. These surfaces formed during eustatic low stands in sea level, have associated palaeosols and exhibit typically covered karst vadose solution effects and alteration phenomena. The lower 'Q' units are strongly influenced by the pre-Pleistocene topography and exhibit pronounced variations in both thickness and discontinuity relief. These effects are quickly abolished however and the higher 'cycles' of carbonate sediment are bounded by virtually flat palaeokarstic surfaces and exhibit fairly constant thicknesses over large areas. Unit Q5, for example, maintains an almost constant thickness of 1.5 m over an area of 400 to 500 sq kms! Complex relationships persist both at the landward margin and in the shelf edge shoals, but future transgressions and regressions affecting the shelf lagoon may be expected to generate successive isopachous units of sediment bounded by relatively flat, laterally extensive palaeokarstic surfaces. Similar Pleistocene surfaces buried beneath a cover of Recent sediments are recorded from shallow water basins in Shark Bay (Hagan and Logan, 1974).

(c) Channels

(i) Description

Incised through palaeokarstic surfaces at several levels throughout the sequence are major channel features and it is within the confines of these that many of the siliciclastic units of the succession occur. These channel sandbodies, which exhibit often complex internal sequences are described more fully in Chapter 4.

Individual channels attain observed widths of up to 700 m although inland mapping demonstrates that locally they may be over a kilometre wide (Fig. 26). Channel depths of up to 25 m are recorded. The sides of the channels vary from gently sloping to notched, terraced and locally vertical (Plates 5, 6b; Figs.25). Cross sections are often markedly asymmetrical (Fig.25). Examination of the channels is usually restricted to two dimensional coastal sections and their three dimensional distribution is largely unknown. An exception is the channel system which contains the Benllech Sandstone. This unit crops out widely inland and mapped relationships indicate that individual channels are single anabranches of a more complex anastomosing pattern (Fig.26). The channels bifurcate and rejoin around upstanding knolls of limestone and these are capped by normal hummocky palaeokarstic surfaces.

Contacts between the limestones and channel-fill siliciclastics are often complex. Conglomeratic units which line the bases of many channels commonly contain or indeed may be dominantly composed of clasts of the adjacent limestones, the 'Limestone Conglomerates' of Greenly (1919) (Plate 6a). A gradation occurs from conglomerate-filled cracks and fissures in the limestones to large detached blocks of limestone within the conglomerate (Plate 6b) fining away from the channel margin. Limestone clasts are generally confined to such contact zones, but tongues of limestone conglomerate extending tens of metres into channel-fill sequences have been observed (Fig.16).

Two channels containing the Benllech Sandstone are exposed at the coast (Section 5.6b). The northern margin of the southern channel (Fig.17; Plate 7) represents the 'anomalous junction at Borth Wen' described by Greenly (1919, p.612). Interbedded trough cross-bedded conglomerates and bioturbated silty shales give way across a narrow gully to bioturbated and flaser bedded sandstones and shales with intercalated limestone breccias and a capping sequence of tilted limestone beds. This capping limestone sequence can be matched to adjacent, but more elevated exposures of the Lower Dinas Beds and apparently represents a slumped mass of limestone. Two slide planes have been identified. Affected sandstone beds display truncation, convolution of internal lamination, drag folds and splaying. The back scar of the slump exhibits sandstone-filled fissures.

(ii) Interpretation

The stratigraphic relationships illustrated in Fig. 25. leave no doubt that the channels were contemporary features of the palaeokarstic surfaces through which they are incised. The depositional environments envisaged for the siliciclastic fills are discussed in detail in Chapter 4. The characteristics of the conglomerates in particular (Section 4.5a) and the anastomosing channel patterns indicate a fluvial origin. The channels therefore demonstrate the fluvial dissection of the limestone platform during regressive periods of the sequence and record the, at times, drastic

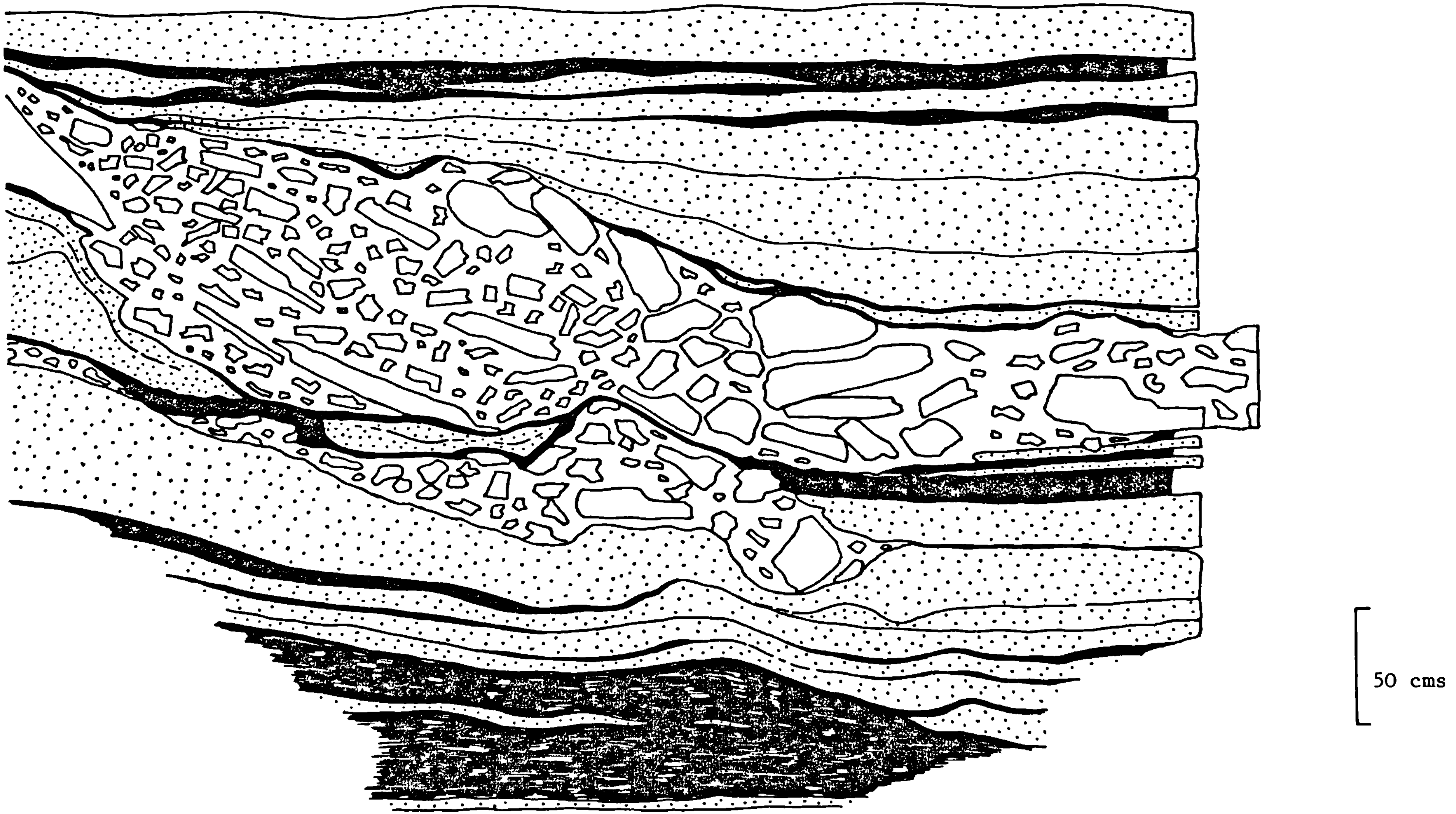


Fig.16. Tongues of limestone clast conglomerate intercalated within sandstones and shales, of the Benllech Sandstone; close to the unexposed southern margin of the northern channel [5201 8407].

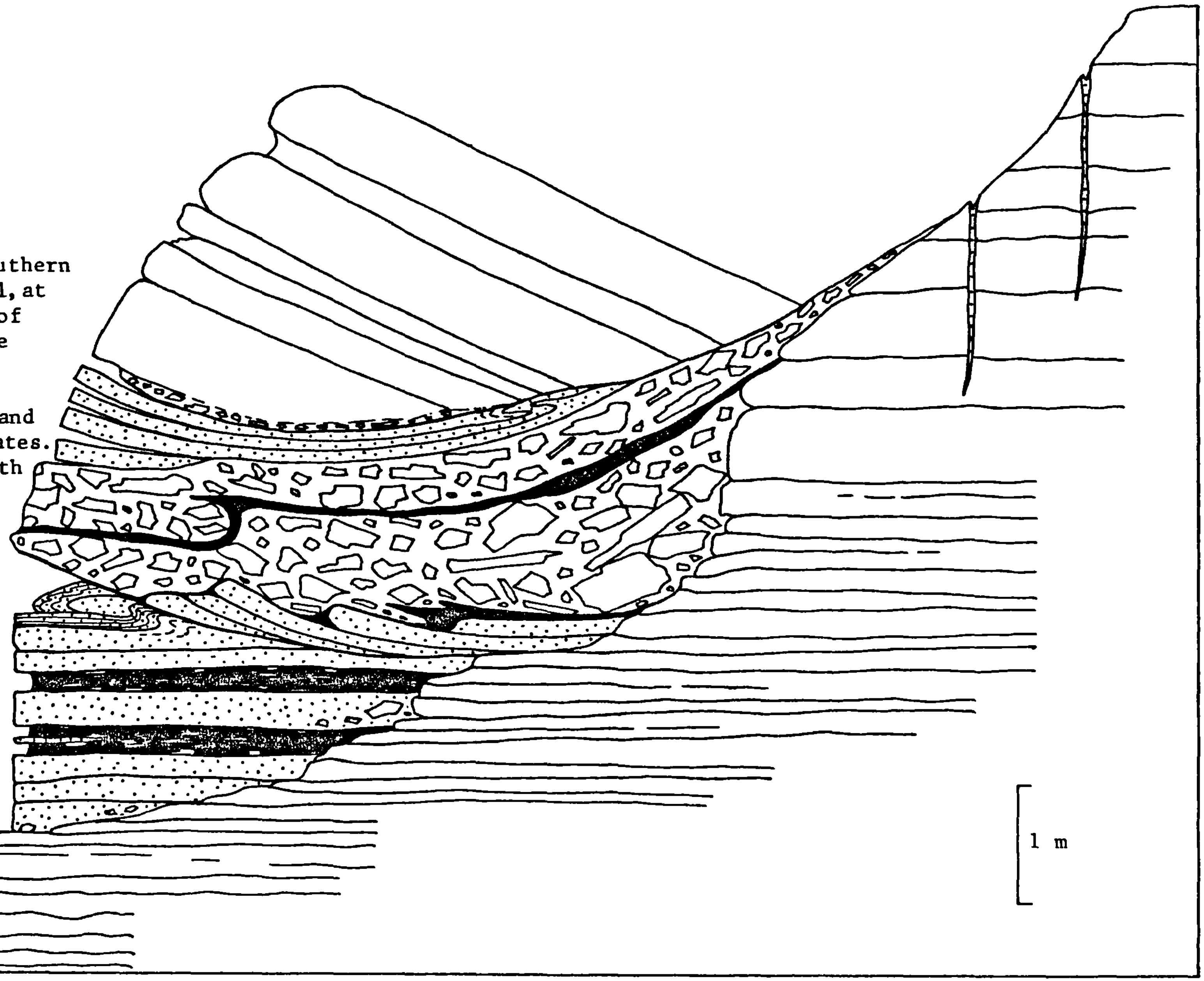


Fig.17 Northern Margin of the southern Benllech Sandstone channel, at Borth Wen : slumped mass of bedded limestones from the adjacent Lower Dinas Beds resting on splayed and contorted sandstone beds and limestone clast conglomerates. A second slipage plane with associated contorted sandstone beds is present beneath the conglomerates.

lowering of base level, up to 25 m, associated with these events.

The following senario outlines the probable sequence of events and processes which operated during the formation of a channel complex. It is specifically for the Benllech Sandstone complex since this best displays the three dimensional properties and the marginal phenomena of the channels (see also Section 4.6). The anastomosing channel pattern was probably established during the intial stages of regression as streams, with sources in the hinterland of older rocks, flowed across the newly exposed, poorly lithified carbonate platform. As regression continued channels became incised in response to the lowered base level, whilst the increasingly indurated nature of the limestones, a result of vadose diagenesis, precluded lateral migration. The channels became 'fixed' and the islands of limestones created by their initial bifurcating pattern became isolated as upstanding knolls within the channel complex (Fig.18).

Siliciclastic sediment derived from the hinterland of older rocks was carried down the channels. Active accumulation of this material however would only have occurred as the gradients of the channels reached equilibrium with the new base level, or indeed as this began to rise in response to the ensuing transgressive episode. Marginal limestone clast breccias are interpreted as scree deposits which accumulated along the channel sides and interfingered with the channel clastics. Conglomerates rich in limestone clasts may represent reworking of these or of limestone masses slumped off the channel sides. It is of course tempting to regard the Borth Wen slump as a contemporary feature of the Benllech Sandstone channel. Certainly the obvious soft

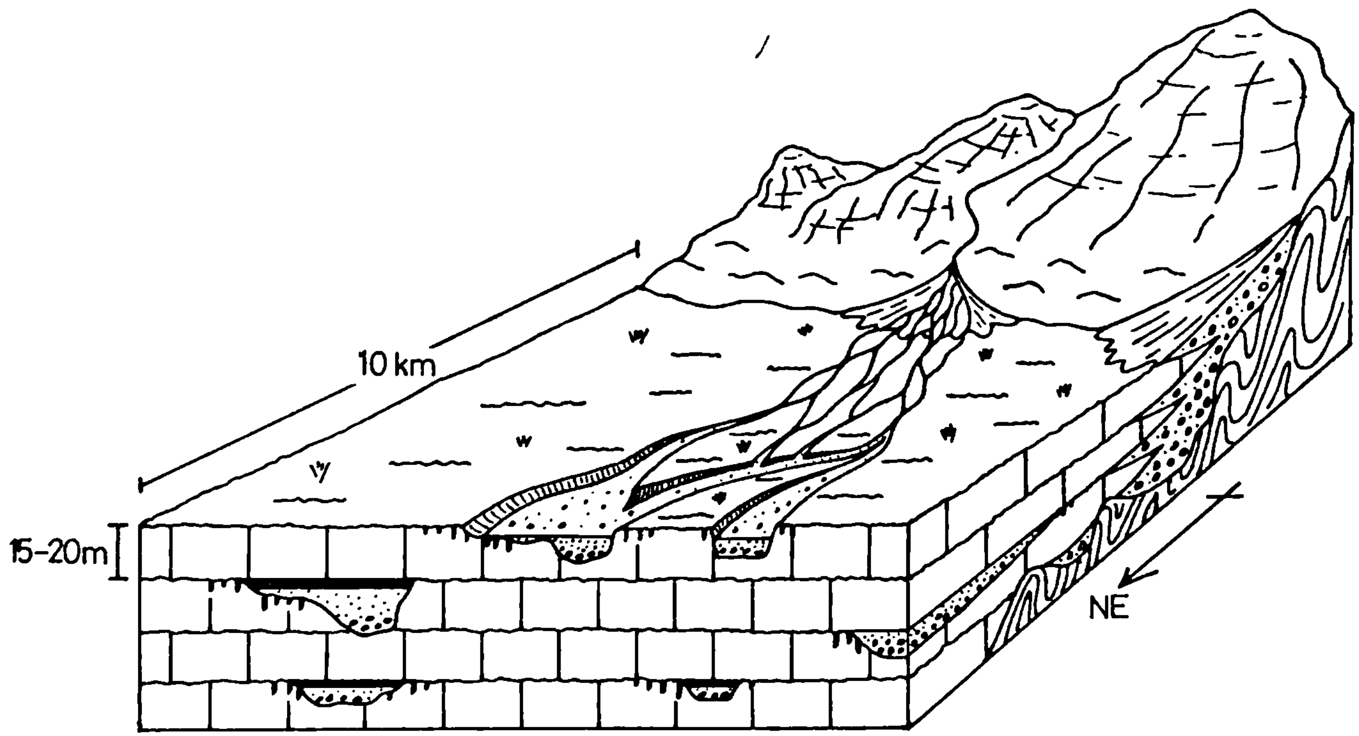


Fig.18 A regressive period during the Anglesey Dinantian : complex anastomosing channel systems are incised through contemporary palaeokarstic surfaces and it is within the confines of these that transportation and deposition of siliciclastic sediment largely takes place.

sediment deformation structures, recorded from the associated sandstones indicate an origin prior to lithification, but this does not exclude early post burial compactional effects and the exact age of movement remains equivocal.

Transgression caused the drowning of the channel system and the establishment of estuarine conditions (Section 4.5a). Rising sea level eventually lead to inundation of the adjacent palaeokarstic surfaces and a return to active carbonate deposition on the shelf.

(d) Sandstone pipe horizons

(i) General Description

Several palaeokarstic surfaces throughout the Anglesey Dinantian exhibit the peculiar sandstone pipes for which the succession is famous (Greenly, 1900 and 1919; Morton, 1901; Challinor and Bates, 1973 and Bates and Davies, 1981). More detailed analysis of these phenomena is provided by Baughen and Walsh (1980) and Walkden and Davies (in prep.).

The pipes are sandstone-filled pits up to 5 m deep and 3 m wide. In several instances they are observed piercing the upper slopes or terraces of channel margins or occur in adjacent palaeokarstic surfaces up to 300 m beyond. They have never been observed piercing normal hummocky palaeokarstic surfaces where there is an associated overlying bentonitic palaeosol and it appears likely, therefore, that the pipes are marginal features of channel complexes.

Pipes vary from deep, narrow and cylindrical to less well defined conical and bowl shaped pits. The deeper pipes generally exhibit smooth, vertical, locally overhanging sides often with concentric ribbing. Bases are not commonly observed, but appear to be smooth, well defined and bowl shaped (Plate 12c). Shallower pits tend to be more irregular with fretted or digitate margins and bases. Cross-sections are commonly circular or slightly ovate, though where pits coalesce (Plate 11b) or display channel-like interconnections (Fig.20) more complicated patterns result. The pipes typically penetrate their host limestones normal to bedding, but occasionally their long axes may be orientated at shallower angles e.g. at Porth yr Aber.

The sandstone fills may extend from the base of overlying sandstone units (eg Plate 15) the 'source sandstones' of Baughen and Walsh. In other cases there is no contiguous sandstone layer and the sediment preserved in the pits forms the only record of siliciclastic deposition at such levels. The fills vary from coarse conglomerate or pebbly sandstone to well sorted, fine-grained siliceous sandstone.

Many sandstone fills exhibit a characteristic form of internal lamination. Where present laminae towards the top of the fill are disposed horizontally or gently concave downwards. Successively deeper laminae become increasingly concave upwards such that the lowest ones are distinctly paraboloid in form and towards the margins of the fill become sub-parallel to the pipe walls. This structure is here termed meniscus-type lamination. A further informative feature of the pipe fills is the behaviour of bedding present within source sandstones. Such bedding geniculates sharply as the fill descends into the pipes and becomes inclined at high angles (Fig.21).

(ii) Locality Details

Twelve sandstone pipe localities have been recorded. Detailed description of many of these is provided by Baughen and

Walsh (1980) and of the Trwyn Dwlban horizon in particular by Walkden and Davies (in prep.). The main features of the localities are outlined below. They are described in ascending stratigraphical order.

Pedolau [5100 8700] : The gently rising eastern slopes of the channel containing the Pedolau Sandstone exhibits a discontinuous veneer of coarse, massive, poorly sorted and reddened quartz pebble conglomerate (Plate 8a). In places this descends into circular, ovate or bulbous shaped pits in the underlying surface (Plate 8b). These are seldom over 50 cms deep and vary from just over 10 cms (Plate 8c) to 3 m in diameter. The contacts between the pits and the host limestones are generally sharp and well defined, but where they impinge into more rubbly or brecciated areas the margins become difuse. No internal structures have been observed in the conglomeratic fills.

Porth Helaeth [5102 8695] : Towards the western side of Porth Helaeth coarse conglomerates of the Helaeth Sandstone impinge on the bevelled upper surface of the Royal Charter Beds (Plate 9a). Recent erosion has revealed broad, dish shaped masses of the conglomerate up to 2 m across preserved within shallow isolated hollows in the limestone surface (Plate 9b). Peculiar depressions within the main body of the conglomerates (Plate 9c) may be related to underlying, unexposed examples of these features.

Moelfre Coastguard [5144 8684] : Sandstone pipes developed in the palaeokarstic surface separating the Upper Helaeth and the Lower Lookout Beds are exposed in a fault-repeated section to the west and north of the Moelfre Coastguard Lookout. At this locality the sandstone pipes are isolated within the palaeokarstic surface, there is no overlying 'source sandstone'. The often complex relationships displayed by the pipes of this horizon are illustrated in Plate 10 and Fig.20. They pierce a 25 to 30 cm thick calcite mudstone bed and then commonly geniculate sharply along its lower contact before penetrating the underlying skeletal grainstone. In other cases there is an abrupt narrowing of pipe diameter. The fills of adjacent pipes are often linked by ill-defined ribbons of sandstone along this bedding surface and in plan these appear to be filling ill-defined channels in the top of the grainstone.

The pipes are filled by light brown, slightly calcareous, pebbly sandstone. No internal structures have been observed. The margins and bases of the pipes are often highly irregular due to the brecciated and fissured nature of the host limestones. In many cases this brecciated marginal zone has been preferentially eroded and the central plugs of sandstone is all that remains. The tops of these sandstone plugs often stand proud of the calcite mudstone bed and overlying limestones 'dome' over them.

Eglwys Siglen [5159 8692] : The sandstone pipes at this locality pierce the palaeokarstic surface which marks Asbian/Brigantian boundary, the contact between the Moelfre and Traeth Bychan Limestone Formations (Section 3). This forms one of the more impressive sandstone pipe exposures, recognised by Greenly (1919) and described in detail by Baughen and Walsh (1980, p.17). The pipes, up to 2.2 m deep, are observed piercing a narrow coastal ledge which represents the exhumed palaeokarstic surface (Plate 11a). The latter authors were able to map the distribution of the pipes within this surface (op. cit. fig.6) and recognised over 160 examples. No 'source sandstone' is present. The fills consist of yellow, pebbly sandstone with a marginal sheath of silty shale. Meniscus type internal lamination is commonly observed. Recent

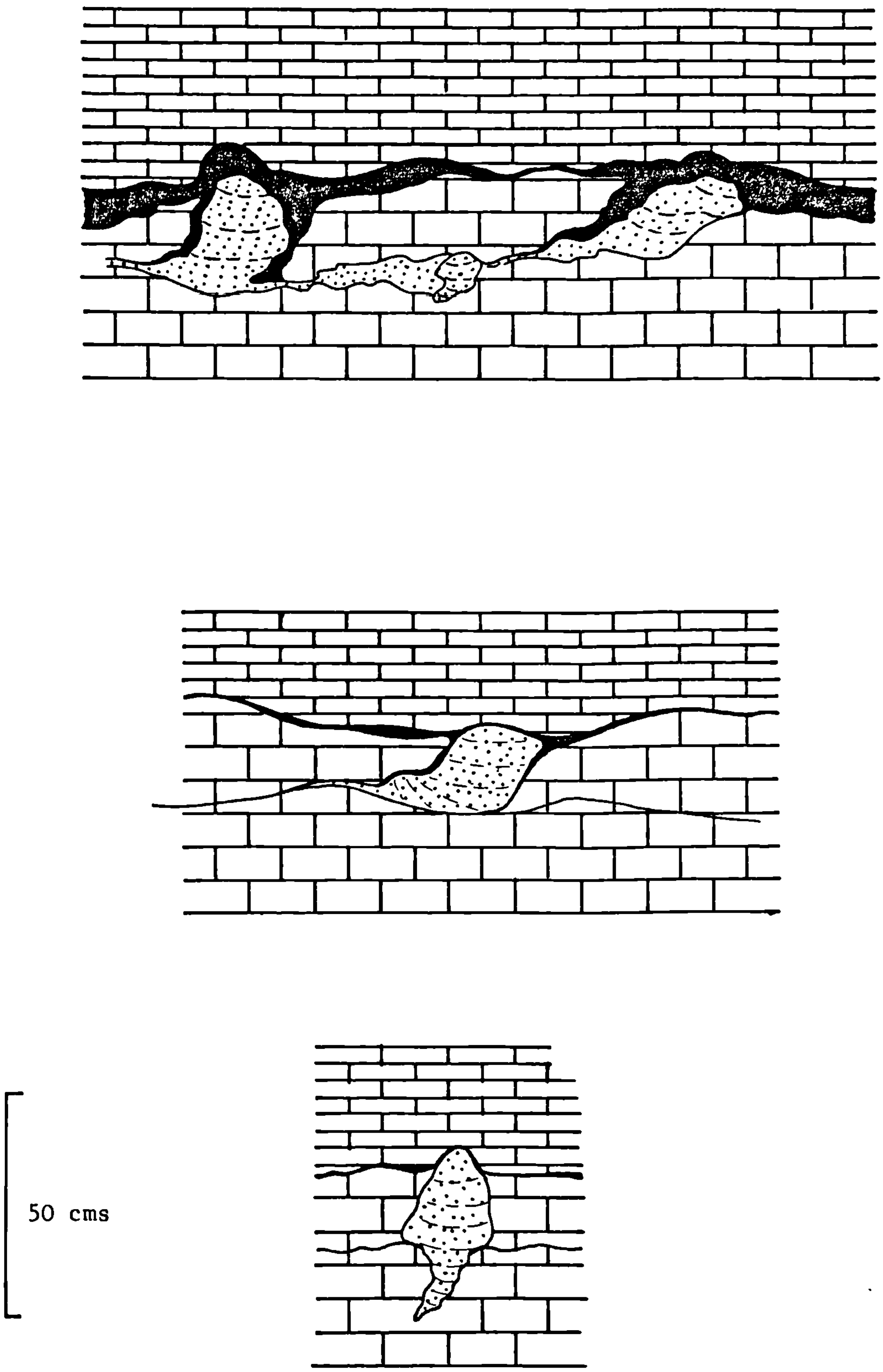


Fig.20 Sandstone pipes at the top of the Upper Helaeth Beds [5144 8684].

marine erosion however has often scoured out the sandstone fills and many of the pipes, cylindrical and funnel shaped, are now empty. The coalescence of pipes, in places several in a line, has lead to irregular bulbous and chain-like cross sections (Plate 11b).

Ynys Moelfre [5187 8689] : Owing to difficulties of access the Ynys Moelfre (Moelfre Island) locality has only been visited once. Two sandstone pipe horizons are present. The upper one is equivalent to that exposed at Eglwys Siglen. The lower example, which has a sandstone source rock, is represented on mainland by a 'normal' unpiped palaeokarstic surface. The exposure is much affected by marine erosion and is heavily encrusted by barnacles and algae. Detailed observation has not been possible.

Porth yr Aber [5144 8571] : The sandstone pipes at Porth yr Aber are illustrated in Fig.21 and Plate 12. Two pipe horizons are present. The lower, most prominent set pierce the top of the Porth yr Aber Beds and are filled by units of the overlying Aber Sandstone. Well defined cylindrical pipes up to 3 m deep and 2.5 m wide penetrate massive bioclastic and oolitic limestones. The pipes have smooth concentricly ribbed walls and sharp bowl-like base (Plate 12c). Overhanging lips are developed in some examples. Cross-sections are circular or nearly so. Few of the pipes are vertical, but instead tend to slope in a northerly direction at angles of 60° to 70°.

The fills descend from the base of the overlying Aber Sandstone. Only the lowest two beds of this sandstone unit are involved in the filling process. The lower of these beds is a coarse, argillaceous sandstone with conspicuous undulations on both its top and base. The upper sandstone bed is finer grained and contains fossil rootlets. These beds are separated by a 2 to 5 cm thick silty shale band.

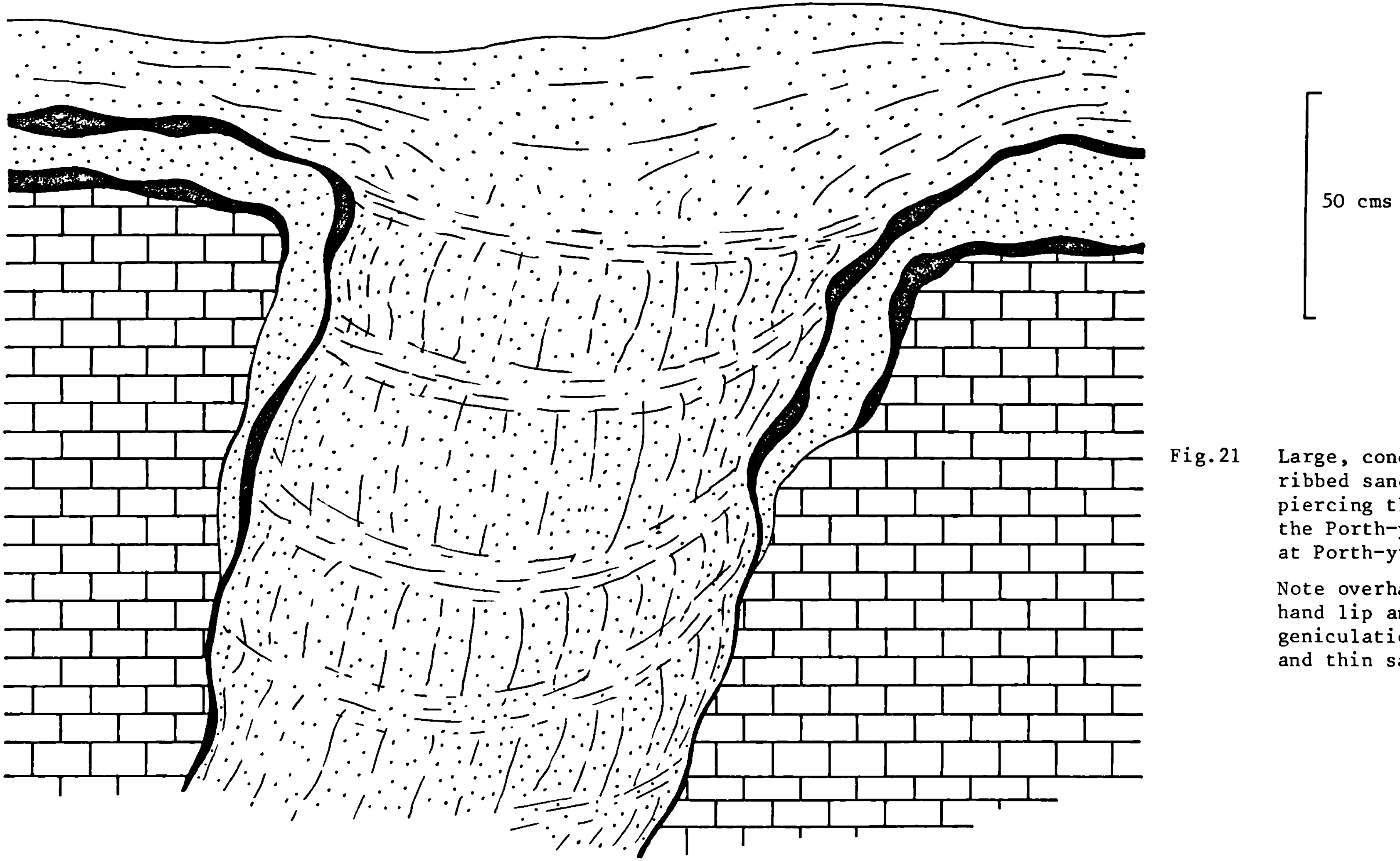


Fig. 21 Large, concentrically ribbed sandstone pipe piercing the top of the Porth-yr-Aber Beds at Porth-yr-Aber.

Note overhanging left hand lip and geniculation of shales and thin sandstone bed

These units geniculate sharply into the pipes and bedding contacts become steeply inclined and in places vertical. The silty shale band is often pinched out. The upper bedding surface of the higher of the two sandstone beds is generally flat or may exhibit a slight doming over the pipes. The succeeding calcareous sandstones are unaffected.

Towards the top of the calcareous upper units of the Aber Sandstone a further set of pipes is developed. Up to 40 cms wide and 30 cms deep these vary from well defined, sharp sided conical forms to highly irregular with fretted and digitate margins. The fill which may show well developed meniscus-type lamination (Plate 12b) descends from the topmost bed of the Aber Sandstone and consists of calcareous pebbly sandstone.

Frigan Quarry [4845 8356] : One of the few inland localities the sandstone pipe horizon exposed in Frigan Quarry equates with the lower of the two sets at Porth yr Aber. The pipes are filled by an overlying coarse, poorly sorted, pebbly sandstone which is the lateral equivalent of the Aber Sandstone at the coast. No upper pipe horizon as at Porth yr Aber is present at this locality.
A series of 5 or 6 pipes are observed (Fig.22a) varying from shallow sharp sided rounded pits up to 20 cms deep and 25 cms wide to highly irregular forms which interdigitate with the host limestones and taper downwards to depths of over 40 cms (Fig.22b).

Figin-fawr [5112 8472] : The palaeokarstic surface between the Upper and Lower Dinas Beds has been exhumed to form a broad inland dip slope on which Figin-fawr Farm is built. Top soil covers much of the feature but bed rock is exposed along the track which leads to the farm. Several circular patches of light brown, fine-grained sandstone up to 1.5 m in diameter are observed and are thought to

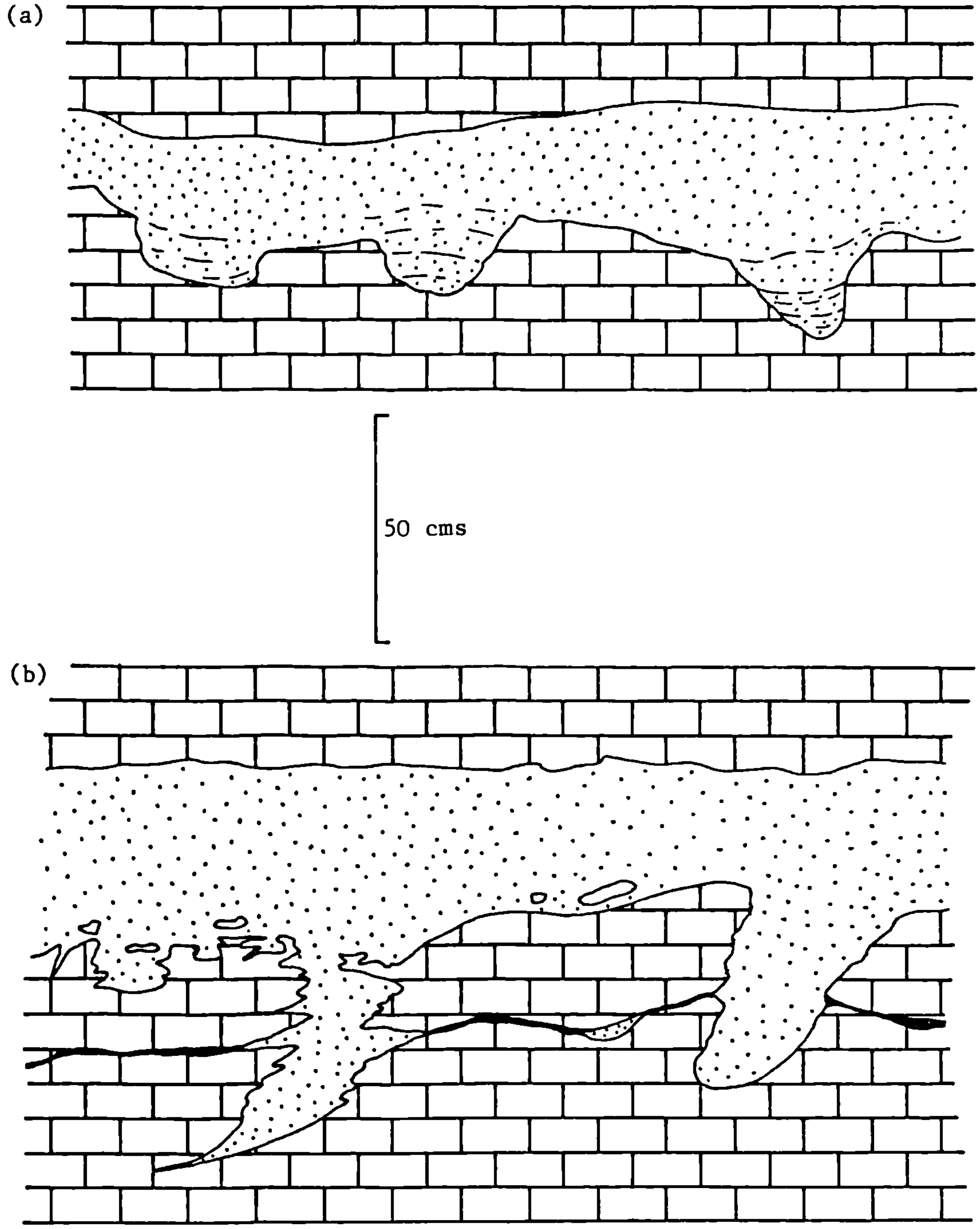


Fig.22 Sandstone pipes exposed in Frigan Quarry [4845 8356] descending from a thin source sandstone which separates limestones of the Porth-yr-Aber Beds (Lower) from the Porth y Rhos Beds (Upper).

represent the tops of cylindrical sandstone pipes. No source sandstone is now exposed. The locality lies 300 m to the north-west of the channels containing the Benllech Sandstone which are also incised through this stratigraphic level.

Huslan [5220 8340] : The sandstone pipes exposed on the broad coastal platform between Huslan and Pen-y-coed occur at the same stratigraphic horizon as those at Figin-fawr i.e. at the contact between the Upper and Lower Dinas Beds. Here, however, the pipes are developed in calcareous sandstones, extensions of the Benllech Sandstone (see Section 5.6b) the channel for which is exposed immediately to the north in the cliffs at Pen-y-coed. The calcareous sandstones, up to 1 m thick, rest on the palaeokarstic surface through which the channel was incised and which defines the top of the Lower Dinas Beds. The Huslan pipes are comparable with the higher of the two horizons at Porth yr Aber, where piping of a similar host rock has been described. The Huslan examples exhibit similar fretted and digitate margins. They tend to be broad shallow features up to 1.5 m wide but seldom over 50 cms deep. The fill is light brown, fine-grained sandstone which also occurs as a discontinuous veneer, up to 15 cms thick, on the upper surface of the calcareous sandstone host.

Trwyn Dwlban [5324 8191] : Perhaps the most famous of the sandstone pipe localities, often visited by student parties and first noted by Henslow (1822). It exhibits well the features and properties of the sandstone pipes and forms the specific subject of a paper, in preparation, by the author and Dr. G.M. Walkden of Aberdeen University. Further detailed description is provided by Baughen and Walsh (1980), but they failed to record some of the more unusual and informative relationships visable at this locality.

The pipes are developed in the palaeokarstic surface which defines the top of the Lower Dwlban Beds and are associated with the channel filling Dwlban Sandstone (Fig.37). The main pipe locality is situated on a terrace-like feature to the south of the main conglomerate filled channel (Plate 13b). A minor tributary channel, lined with conglomerate, is observed incised into this feature (Plate 13a) The thick conglomerate deposit of the channels are reduced to a thin (up to 10 cm thick) conglomeratic veneer which adheres to much of the piped surface. This veneer is pierced by and therefore predates the main piping episode (see below). Units involved in filling the main pipe set comprise a silty shale bed with plant debris up to 20 cms thick, which overlies the conglomerate veneer, and a 20 to 30 cm thick buff, medium grained and conspicuously burrowed sst which comprises the main 'source sandstone'. The succeeding thick bed of silty shale, overlain by limestones of the Upper Dwlban Beds, is not involved in the filling process (for more detailed analysis of the Dwlban Sandstone see Section 4.6d).

Over 150 pipes have been recognised and the extensive nature of the exposure allowed Baughen and Walsh (op. cit. fig.3) to map their distribution. They appear to be evenly distributed across the surface and reflect no structural control by the host limestones. The pipes are typically cylindrical with smooth sides, rounded bases and circular cross sections. They are up to 3 m deep and 1.5 m wide piercing their host limestones normal to bedding.

The fills generally of the overlying buff sandstone descend abruptly into the pipes whilst the intervening silty shale bed geniculates sharply forming an attenuated marginal sheath around the central sandstone plug. Bedding contacts become vertical. Several of the sandstone fills exhibit well developed meniscus-type lamination (Plate 12c). To the south of the locality the upper surface of the buff source sandstone bed is extensively exposed. Conspicuous dome-like prominances indicate the position of underlying pipes (Baughen and Walsh, fig.3).

About 15 pipes have been observed which instead of the normal buff sandstone are filled by yellow-brown, fine grained conglomerate and which have no marginal shale lining. These represent an earlier generation of pipes and are in places truncated by the more common shale lined, buff sandstone filled set (Plate 14). Such relationships clearly indicate the lithified nature of these earlier conglomeratic fills prior to main phase of piping. These earlier fills are moreover often overlain by the conglomeratic veneer described above the latter distinguished by the abundance of limestone clasts it contains; indeed this veneer often thickens into moat-like depressions around the margins of these pipes. These early structures must, therefore, also predate conglomerate deposition within the Dwlban channels.

<u>Dwlban Quarry</u> [5322 8172] : Two sandstone pipes are exposed at this locality and are the largest recorded from the succession, the western one measuring up to 5 m deep and 3 m wide (Plate 15). They are developed in the next palaeokarst above the Trwyn Dwlban locality, at the top of the Upper Dwlban Beds. The pipes are filled from the overlying St. David's Sandstone, which for the short distance it is exposed appears to have a sheet-like geometry.

The eastern of the two pipes is rounded and bulbous in form whilst the deeper western pipe is more funnel-shaped. As the overlying sandstones descend into the pipes bedding typically bends sharply and intercalated shale beds become pinched and distorted. The main body of the St. David's Sandstone descends only shallowly into the pipes (Plate 15a) and demonstrates that much of the fill was derived from source sandstones no longer present beyond the confines of the pipes.

Parc Trwyn-du [6373 8140] : This, the only sandstone pipe horizon recorded from the Penmon Area, is located some 400 m to the west of Trwyn Du Lighthouse. Here coastal erosion has provided extensive exposure of a piped palaeokarstic surface dipping at 20° to the north-east. This horizon separates Penmon minor cycles TB6 and TB7 which are correlated with the Lower and Upper Dinas Beds of the Principal Area (Section 3.4e). The Parc Trwyn-du piped surface therefore equates with those described above from Figin-fawr and Huslan. Indeed the pipes are closely comparable with those at Huslan since they penetrate a calcareous sandstone bed which rests on the true palaeokarstic surface. At Parc Trwyn-du, however, where there is no overlying source sandstone, the pebbly sandstone fills often pass right through the calcareous sandstone bed and pierce the underlying limestones.

Baughen and Walsh (op. cit. fig.8) have mapped the distribution of sandstone pipes, of which there are over 200, across this surface. There appears to be no discernible pattern. The pipes are generally small, up to 1 m deep and 60 cms wide, and are mainly cylindrical in form. The fills have often been removed by marine erosion. The apparent ovate cross-sections of many of the pipes were thought by Baughen and Walsh to be a primary feature. In fact the elongation of many of the pipes is along fault related tension gashes and the ovate cross-sections may therefore be related to later tectonic modification.

(iii) Interpretation

The various mechanisms put forward in explanation of the

Anglesey pipe horizons have been reviewed and discussed by Baughen and Walsh (1980) and Walkden and Davies (in prep.). The absence of disturbed bedding within the host limestones, the truncation of fossils and indeed of an earlier genations of pipes at Trwyn Dwlban indicate the erosional origin of the pipes. The injection and loading mechanisms suggested by Hobbs (1907) and Greenly (1919) are clearly untenable. The marginal setting of several of the pipe localities to fluvially incised channel complexes (Section 3.2c) instantly suggests a pothole type origin. However, the geometry of the fill material, the absence of coarser pebbles in the bases, and the occasional fretted and digitate nature of the sides all argue against such a mechanical origin. The shallow pits at Porth Helaeth, filled and overlain by fluvial conglomerates (Section 4.6a), may indeed have been formed by current scour, yet the depressions present within the conglomerates (Plate 9c) indicate the continued deepening of these features after burial. These and other pipes associated with channel margins may have been initiated by mechanical processes but their present form is clearly the result of some other mechanism.

The geometry of the sandstone fills and their relationship to overlying source rocks provide the clearest evidence as to the origin of the pipes. The geniculation of bedding into the pipes often becoming vertical and the presence of oversteepened meniscus-type lamination indicates concomitant deepening of the pipes and subsidence of the fill. The only feasible way of producing these effects is by dissolution of the host limestones beneath an unlithified sand cover.

Since the sandstone pipes are features of palaeokarstic surfaces and appear related to periods of lowered base level the most obvious mode of formation is dissolution by downward percolating vadose groundwaters. Baughen and Walsh draw attention to the gross similarities of the Anglesey pipes to solutional pipes formed in this way in the Chalk. These form at the interface between the Chalk and overlying Tertiary sediments. Meniscoid fills are developed as the cover rocks slump into the growing pipes.

Further recent analogues have been observed developed in Tertiary calcareous siltstones in north-east Spain by Dr. C.F. Klappa (pers. comm.). Plate 16a,b are his photographs of these phenomena. Cylindrical and conical pits pierce exposed terrace-like surfaces adjacent to dry river valleys, a setting comparable with the channel margins of many Anglesey pipes. The pits are filled by insoluble residues derived from the solution of the calcareous siltstones and this acts as soil for isolated clumps of vegetation. These fills and vegetation they support aid in the solution and enlargement of their host pits both by adding organic acids to the infrequent rain water and by slowing down the rate of water loss due to evaporation or soak-away. Clearly the fills descend commensurately with pit deepening.

Solutional pipes described by West (1973) from calcareous raised beach deposits in Cornwall are closely comparable with those developed in calcareous sandstones at Porth yr Aber, Huslan and Parc Trwyn-du. These latter sandstones, Lithofacies C of the siliciclastic scheme, are interpreted as beach or upper shoreface deposits (see Section 4.5a). The development of solutional and minor emergent phenomena within such units suggest that a raised beach analogue may be very close indeed. The Cornish pipes form by solutional decalcification of the sand. Non-calcareous sandstone fills in both these recent and the Anglesey pipes require no source sandstone, the fills merely represent the residual quartz sand left after decalcification. Such intimate relationships between host and fill are reflected in the fretted and digitate margins of the Anglesey pipes.

Cylindrical solution pipes associated with palaeokarstic surfaces in Pleistocene limestones in Florida (Perkins, 1977) are again a product of vadose dissolution.

3.3 BENTONITIC PALAEOSOLS

(a) Description

Most palaeokarstic surfaces are at some point overlain by distinctive clay seams. These are readily removed by recent erosion and give rise to prominent rentrants in the coastal cliff sections. The seams envelop underlying palaeokarstic surfaces thickening into the hollows and thinning or even pinching out completely against the upstanding hummocks (Plate 4c). Thicknesses clearly vary markedly from up to 2.25 m in some of the deeper pits to zero; average values are around 10 to 20 cms. When fresh the clays vary from firm, but pliable to being quite hard with a blocky fracture. They are generally mottled pale green and grey in colour. Branching carbonaceous streaks are common and may display bright purple to mauve halos imparting a vividly variegated appearance (Plate 17a). Brick red varieties are observed in the bases of some of the deeper hollows or in the lower portions of thicker seams (Plate 17b Thin streaks of coal and scattered plant debris have been observed in several examples. Silty and micaceous varieties also occur. Listric surfaces are common and euhedral crystals of pyrite or chalcopyrite are ubiquitous. On weathering, oxidation of the pyrite results in conspicuous ochreous yellow and orange hues and the clays become soft and sticky or crumbly in texture (Plate 18a).

Contacts between clay seams and underlying strata reflect the commonly smoother nature of the palaeokarstic surfaces and are generally sharp. Deep rubbly profiles are developed locally, however, and here the junction is necessarily illdefined with lenticles and stringers of clay extending 1 to 2 m below difuse palaeokarstic surfaces (Plate 18b). Upper bedding contacts are again either sharp with the base of overlying limestone units flat to gently undulating, or display an intimate mixing of the two lithologies.

(b) Palaeosol Geochemistry

A detailed geochemical study of clay seams overlying Dinantian palaeokarstic surfaces in North Wales has been undertaken by Somerville (1977). This follows previous studies by Walkden (1972) on clay 'wayboards' in the Dinantian of Derbyshire. In light of this in-depth treatment of the subject a similar investigation for the Anglesey examples was felt unwarranted. Only two samples of Anglesey clay seams have been analysed using standard X-ray diffraction techniques simply to establish the similarities with those described by Walkden and Somerville.

Sample 1 came from a conspicuous clay seam exposed in Flagstaff Quarry [6343 8083] 8 m below the top of the Careg-Onen Limestone Formation and is of Asbian age. Sample 2, of Brigantian age, was collected from the Traeth Bychan Limestone Formation from a seam exposed at the top of the Lower Morcyn Beds in Caravan Quarry, Traeth Bychan [5131 8489]. Care was taken to obtain as fresh samples as possible.

The samples were prepared for X-ray defraction analysis using methods outlined by Walkden (p.148). The samples were analysed first in an untreated state, secondly following glycolation and thirdly after heating to 300°C for three hours. The resulting X-raydiffractograms are given in Fig. 23. The traces compare closely with those figured by Walkden (fig.2) from the Derbyshire clays. In those for the untreated samples the illite peaks around the 10 Å mark display either pronounced assymmetry (sample 1) or displacement (sample 2) both indicating interstratification with another clay mineral i.e. the presence of mixed-layer clays. For sample 1 glycolation resulted in a slight shift (9.8 Å to 9.5 Å) and a sharpening of the peak, whilst heating obliterated these effects and produced a well defined illite peak at 10 Å. In sample 2 the effects of glycolation are less marked although a slight sharpening is observed. Heating once again gives a well developed 10 Å illite peak.

These results demonstrate that smectite is the interstratified mineral. The downward shift and/or sharpening of the peaks following glycolation of the samples results from expansion of the smectite lattice, whilst heat treatment collapses this lattice to an illite-like material giving a strong combined peak at the illite 10 Å position.

The Dinantian clays analysed by Somerville (1977) from the Llangollen and Mold districts of North Wales were similarly smectite rich. His more detailed geochemical investigations also showed these deposits to be rich in potassium (4-8% K_2O) and to have high Zr and Rb concentrations. The Anglesey clays differ from those described by Walkden and Somerville in containing larger amounts of chlorite and quartz.

(c) Interpretation

Walkden and Somerville emphasize the similarities of these Dinantian clays to potassium-rich clays known throughout the Palaeozoic and termed K-bentonites. Illite/smectite mixed layering is a characteristic feature of these clays which are believed to result

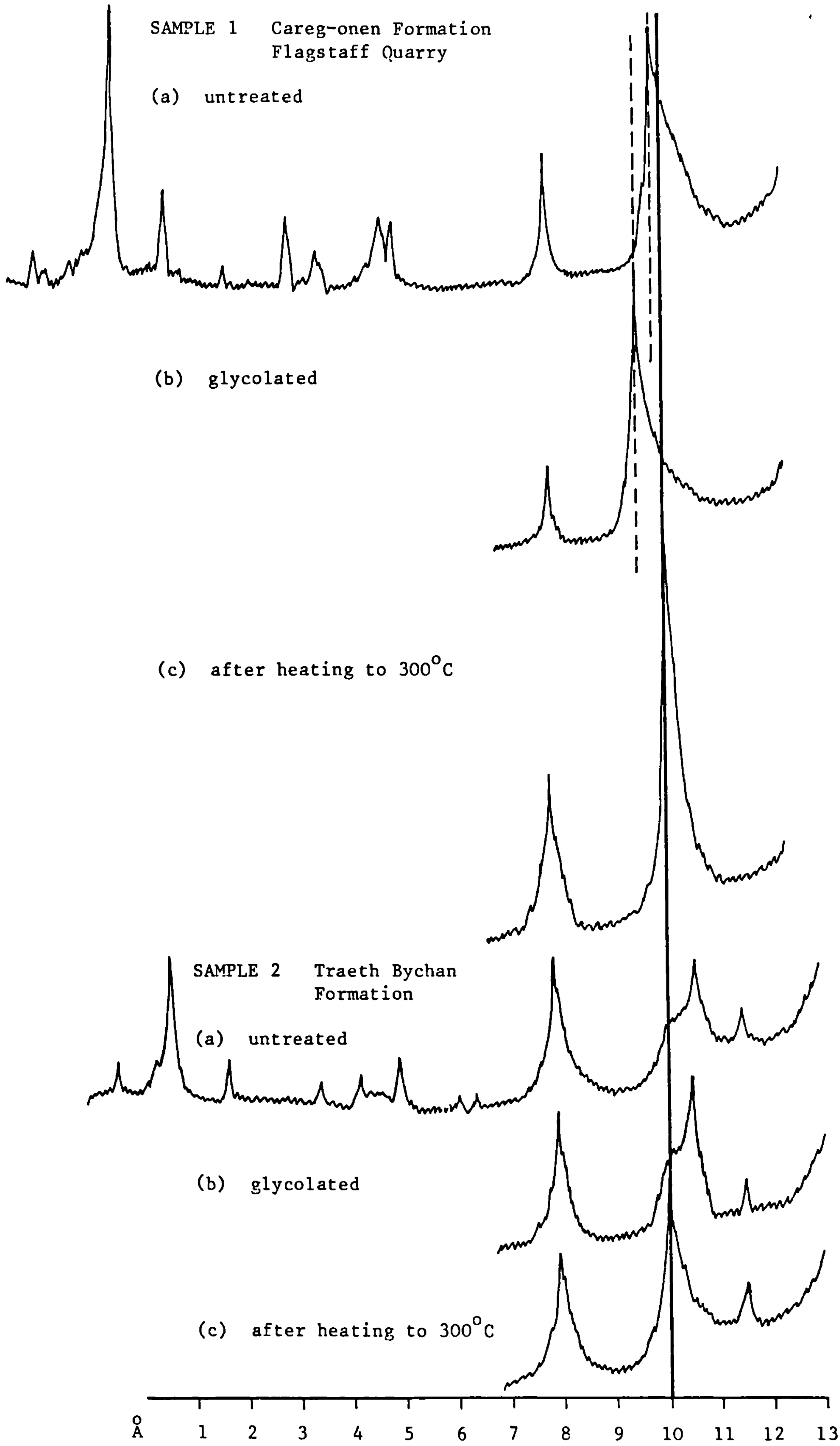


Fig.23 X-ray defractograms of palaeosols

from the diagenetic alteration of wind blown volcanic ash material (see Walkden, 1972, for fuller discussion and references).

Branching carbonaceous filaments present in many clays are interpreted as fossil rootlets and with thin streaks of coal demonstrate the subaerial setting in which these pyroclastic fines were deposited. They accumulated on emergent limestone surfaces and, acting as unstructured, azonal soils, were colonised by surface vegetation. The clay seams are thus interpreted as bentonitic palaeosols. The higher proportionsof quartz and chlorite in the Anglesey palaeosols, as compared with those described by Somerville (1977) and Walkden (1972), are thought to reflect the proximity to an exposed hinterland of older basement rocks in which chlorite schists formed an important element.

Solution beneath these soil layers led to the characteristic covered karst morphologies displayed by the Dinantian palaeokarstic surfaces (Section 3.2b). Soils overlying analogous Pleistocene surfaces in Bermuda (Ruhe et al, 1961) and Barbados (Harrison, 1977) are similarly of volcanic affinity, and are locally rich in montmorillonite, one of the smectite group of clay minerals. Nor perhaps is this similarity merely coincidental. Harrison has discussed the effects of montmorillonitic soils on karst development. Such soils retard the rate of rain-water infiltration through to the underlying limestone bedrock and limit the effects of vadose dissolution. Retention of water in the soil cover concentrates dissolution at the soil/rock interface and leads to the formation of the characteristic hummocky karstic surfaces rather than subterrainean cave systems.

The distribution of the montmorillonitic soils on Barbados is itself a function of a complex interplay of climatic, topographic and ultimately tectonic effects (Harrison, 1977). The montmorillonitic mineralogy appears characteristic of soils in areas of relatively low relief, of low and/or seasonal rainfall and of high evaporation. In contrast upland areas, which experience a more humid climate, are characterised by kaolinitic soils. The more permeable nature of these soils combined with the higher rainfall lead to a dominantly solutional regime and the formation of cavernous or pinnacle karst landforms (Sweeting, 1972). Clearly the former setting appears more pertinent to Dinantian palaeokarstic horizons in Anglesey and indeed across North Wales (Somerville,1979a,c) and Derbyshire (Walkden, 1972).

The brick red clays recognised on Anglesey have not been analysed separately, but may have a different origin from the bentonitic mottled grey and green varieties which invariably overlie them. They are reminiscent of red terra rossa clays which form an important element of some of the Bermuda palaeosols (Ruhe et al, 1961). They are characteristic of karst terrains and form as localised surface accumulations of insoluble material following limestone dissolution (Bridges, 1970). The basal position of the red clays within the thicker Anglesey seams would support such an origin. Thus dissolution beneath a bentonitic ash soil may generate terra rossa-type insoluble residues which accumulate as a basal phase of the soil profile. Alternatively such terra rossa clays may have accumulated within broad depressions in the emergent limestone surface prior to the deposition of the bentonitic ash.

An obvious question is whether the vulcanism which supplied the material for bentonitic palaeosols and the minor cyclicity they help to define are related? Were the tectonic forces responsible for volcanic eruption also the cause of the minor cyclicity or is it simply that prolonged emergence allowed the accumulation of wind blown volcanic material which was always present in the atmosphere? These problems are considered morefully in Chapter 6 and are bound up with the more fundamental question of whether the minor cyclicity is indeed generated by tectonism or is caused by eustatic changes in sea level? From the internal evidence of the Anglesey sequence the volcanic aspects cannot be resolved. The nearest known volcanic centres were in Derbyshire and the Isle of Man. Certainly in the former area some important extrusive events coincide with minor cycle boundaries (Walkden, 1977, but see also the discussion on this paper).

3.4 ALTERATION PHENOMENA OF THE HOST LIMESTONES

(a) Introduction

The alteration phenomena displayed by the host limestones within palaeokarstic profiles are complex and diverse. They are confined to strata immediately below palaeokarstic surfaces and to rubbly blocks within overlying palaeosols. Eroded lithoclasts of altered limestone may occur in the basal beds of succeeding minor cycles. The term alteration is used in its widest sense to include the effects not only of in situ replacement (micritisation and recrystallisation), but also of cementation and brecciation. These processes operated side-by-side and sequentially to give complex overprint textures.

From the above stratigraphic relationships it is clear that the alteration phenomena were contemporary effects of palaeokarst formation. They represent the results of addition and redistribution of $CaCO_3$ within the upper parts of emergent limestone strata where vadose dissolution was a potentially active process. In many, if not all, cases the alteration processes operated beneath a bentonitic soil cover. This is the environment of caliche or calcrete formation and the textures and fabrics outlined below compare closely with such deposits described from both Recent and other ancient limestone sequences e.g. James (1972), Reed (1974), Walls et al (1975), Perkins (1977), Harrison (1977) and Harrison and Steinen (1978).

Modern calcretes form in tropical and subtropical areas where there is a favourable balance between rainfall and evaporation. Too much rainfall results in wholesale dissolution of the limestones, too little in the formation of only a superficial case-hardened crust (Harrison, 1977). In general calcretes characterise areas where there is a net deficiency of moisture such as arid or semi-arid regions, but may also be developed under more humid climatic regions, where there is a pronounced dry season. Under these latter climatic conditions it seems likely that periods of more pronounced solution and karst formation will alternate with periods of active calcrete deposition. The overall solutional morphology of the palaeokarstic surfaces suggests that such a seasonal climatic regime operated during the Dinantian on Anglesey.

Harrison (1977) has discussed the role of a soil cover on caliche development in the Pleistocene limestones of Bermuda. Soils are important for two main reasons. Firstly, they retard the rate at which rain water passes into the underlying limestones and therefore promote caliche formation rather than wholesale dissolution. Secondly, they may have a profound effect on the chemistry of the waters which perculate through them. Partial pressures of CO_2 are increased and complex organic compounds acquired. The enhanced chemical potency of these waters is probably of critical importance for the complex transformations involved in caliche formation to take place (see also Kahle, 1977). Not all caliche deposits are formed beneath soils, however, and certain types of laminated crust in particular may be developed on exposed limestone surfaces (Harrison, 1977; Multer and Hoffmeister, 1968).

In discussion of caliche (calcrete) deposits much attention is given to the source of $CaCO_3$. This is largely because many of the first recognised caliche deposits both ancient and modern are developed in siliciclastic sequences where the source of $CaCO_3$ is indeed a problem (e.g. Reeves,1970). In carbonate sequences, however, the major source is quite evidently the host limestones themselves. Caliche deposits in limestones result from the localised redistribution of $CaCO_3$ rather than requiring a large extraneous source. Solution and reprecipitation is achieved by downward perculating meteoric waters, but also by waters moving upwards by capillary action in response to surface evaporation. Artesian movements of ground water are also important. Additional sources of $CaCO_3$ include wind blown dust, sea spray (James, 1972), base rich clays within overlying soils (Harrison, 1977) and the calcareous tests of terrestrial shelly fauna (Multer and Hoffmeister, 1968).

(b) Palaeokarstic profiles

The various alteration phenomena which characterise palaeokarstic profiles combined with the morphological properties discussed above and the thickness or absence of palaeosols produce a varied suite of palaeokarstic profiles (Fig.24). Profile development is variable not only from palaeokarst to palaeokarst but also laterally along a single horizon with complex profiles passing rapidly i.e. < 10 m into the more simple profile types. An assessment of profile maturity is therefore rendered almost meaningless.

Simple profiles may comprise a single laminated micritic crust veneering a hummocky palaeokarstic surface, overlain by palaeosol and with rhizoliths within the underlying host limestones. Locally the palaeosol may be lost and overlying limestones rest directly on the crust veneer, whilst with subsequent pressure solution and loss of crust material only the rhizoliths within the host limestones may remain to evidence emergence.

In more complex profiles host limestones, with calcretised grains, pass upwards into calcrete ooid grainstones and it is these which

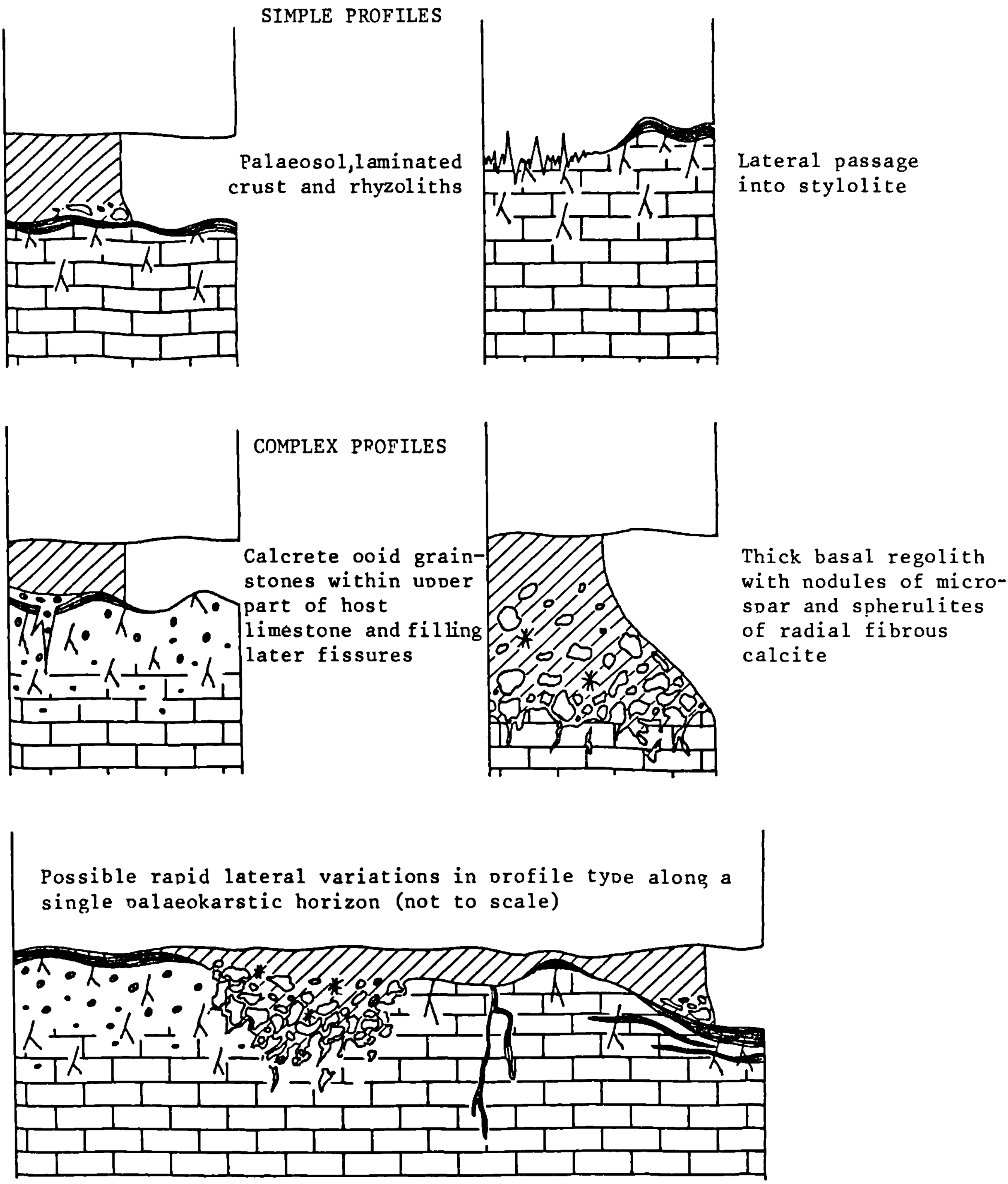


Fig.24 Palaeokarstic profiles

surface crusts veneer. The crusts may in turn be overlain by later generations of calcrete ooids and these may also fill cracks and fissures which penetrate the earlier deposit. Rhizoliths occur throughout; palaeosols may or may not be present. Such ordered profiles may locally be obscured by the effects of penecontemporaneous brecciation. Blocks of both host and altered limestone may be incorporated within overlying palaeosols forming thick basal regolith Elsewhere palaeosol material infiltrates deep within the profile forming extensive rubbly zones in which spherulites and 'chalky' nodules of microspar abound.

Space does not allow all the various alteration phenomena displayed within these profiles to be documented herein. Only the most important textures and fabrics are discussed below and comprise: laminated crusts, rhizoliths, calcrete ooids and some of the effects of recrystallisation, brecciation and cementation.

Other pedogenic features recognised but not described in detail include: calcrete pelloids, comparable with the 'crumb aggregates' of Braithwaite (1975); calcareous glaebules (Brewer, 1964) and hardpan-like accumulations of calcite within palaeosols; various chambered and cellular structures (Plate 38) within the calcretised limestones (c.f. Adams, 1980); geopetal structures and a host of subtle effects related to recrystallisation and cementation.

(c) Laminated crusts

(i) Description

The limestones below palaeokarstic surfaces commonly exhibit a surficial crust of laminated and banded micritic limestone (Plate 11). These crusts may either act as a veneer and faithfully follow the surface irregularities, or may occur as eroded remnants preserved on upstanding hummocks only. Crust thickness vary from an observed maximum of 15 cms to 0.5 mm, the latter occurring as rind-like linings, three to four laminae thick, to fissures within host limestones (Plate 23b). Irregular anastomosing stringers and sheets of crust material up to 2 cms thick also occur within host limestones several centimetres below palaeokarstic surfaces and these may or may not be connected to a surface crust (Plate 21).

Cut blocks of crust material reveal complex fabrics. Fine, highly irregular contorted and crinkly laminae are displayed and are made obvious at this scale primarily due to alternations of colour varying from dark brown, almost black, through a gradation of paler hues to cream. This lamination is commonly superimposed on a crude and coarser banding, up to 2 cm thick, which at first glance is again due to alternations of predominantly dark and light material (Plate 20). Closer inspection however reveals a variety of textural differences between the bands. Some are composed entirely of tightly packed, conformable laminae, others are conspicuous for the abundance of spar filled rhizoliths (see below, Section 3.4d) and/or flattened spar filled fenestrae that they contain. Some bands exhibit a pelloidal texture and appear more detrital in origin, whilst yet others are composed of massive micrite, though seldom without some vague lamination. Brecciated zones also occur. Layers and lenses of calcrete ooids or unaltered skeletal debris have been recorded.
Any one textural type may dominate within a single crust

to the exclusion of the others whilst erosive discontinuities between bands have also been observed (Plate 22).

In thin section the finer lamination within the bands is also observed to reflect similar textural variations, but developed on a microscopic scale and highly complex (Plate 24).
Dense, dark brown cryptocrystalline and micritic laminae alternate with thin laminae of radial fibrous calcite or inclusion rich microspar. These are intercalated with lighter coloured fenestral rich, clotted or pelloidal layers. Intricate spar filled vesicular or cellular structures abound on several scales. These are defined by fine micritic walls or septa and often appear flattened. The finer textures are disrupted in places by the larger spar filled fenestrae and rhizoliths observed in hand specimen. Where the latter are particularly abundant a spongiostrome texture is developed (Plate 26b).

Rare euhedral quartz crystals have been recorded within crusts, whilst some original void spaces are filled by chalcedonic silica.

Laminated crusts rarely overlie unaltered host limestones but generally rest on material which has suffered some micritisation and coating of grains and which is often infested by rhizoliths. The contact is commonly sharp, though often highly irregular particularly where the host limestones have suffered brecciation prior to crust growth. Lamination within the crusts closely follows these underlying irregularities the inherited effects of which may persist for up to 2 cms into crust profiles. The basal 2 or 3 laminae of many crusts are often conspicuously darker and denser than succeeding layers being virtually opaque in thin section. The rind-like linings of fissures are composed of laminae of thin type which are reminiscent of the microcrystalline rinds of Multer and

Hoffmeister (1968) (Plate 23).

(ii) Interpretation

The laminated crusts display in microcosm many of the alteration phenomena which characterise palaeokarstic profiles. Many of the textures and fabrics which form part of the crusts e.g. rhizoliths, brecciation etc. are discussed more fully under their separate headings. The following discussion is concerned therefore not so much with the origins of these component features as with the origin of the lamination and banding they define.

Over the past few years there has been a proliferation of papers describing Recent features termed laminar calcretes or laminated crusts, and it is apparent that similar structures are produced by different processes, in different settings, not all of pedogenic affinity. There appear to be broadly three modes of origin:

1. Direct organic control as in the lichen stromatolites of Klappa (1979) or the desert stromatolites of Krumbein and Giele (1979) both of which develop by accretion on exposed rock surfaces.
2. By in situ replacement of host limestones as suggested by James (1972) and Kahle (1977).
3. By essentially inorganic precipitation and accretionary growth on lithified bedrock either beneath a soil cover or on exposed surfaces (Multer and Hoffmeister, 1968; Read, 1974 and Harrison, 1977), the latter also including the coniatolites of Purser and Loreau (1973).

Given the variety of textures and fabrics displayed by the Anglesey crusts even within a single profile it is possible that several of the above mechanisms were operative in their formation.

The lamination within the Anglesey crusts superficially resembles stromatolitic layering a similarity strengthened by the association with regressive (emergent) phases of minor cycles. Laminar calcretes are often confused with algal stromatolites and the establishment of criteria by which they may be distinguished has been of some importance (Read, 1976). The overall stratigraphic context of the Anglesey crusts i.e. veneering hummocky palaeokarstic surfaces and overlain by bentonitic palaeosols, clearly indicates formation in an elevated terrestrial setting, within the realm of vadose diagenesis rather than on supratidal flats. Recently, however, direct organic control and the term stromatolite have been resurrected in the context of laminar calcretes. Krumbein and Giele (1979) have described modern laminated crusts from California and Israel which are formed by the sediment trapping and calcification of cyanobacteria colonies and which they call 'desert stromatolites'. Klappa (1979) has recognised laminated deposits forming at the present day on exposed calcrete hardplans in the western Mediterranean which result from the calcification of successive growths of saxicolous (rock substrate) lichens. He has termed these 'lichen stromatolites' and has shown that many of the features observed in ancient laminar calcretes, including those on Anglesey, may be interpreted with reference to such a model. "Thus, the organic-rich, dark brown cryptocrystalline laminae are interpreted as calcified lichen thalli; colour variations reflect the distribution of organic matter and iron hydroxides, the wrinkled nature of the humus-rich laminae and the fenestral voids are attributed to desiccation and shrinkage of lichen tissues; calcite spherulites and clotted micrite textures are believed to be calcified cells of coccoid algal bodies and aggregates of decomposed organic detritus respectively . . .". As Wright (1981) has observed "the wheel has turned full circle from a time when stromatolites were re-interpreted as calcrete crusts to the present when some calcrete crusts are being re-interpreted as subaerial stromatolites". Both Klappa and Krumbein and Giele illustrate suits of detailed structures, many only visible using scanning electron microscopy, which characterise their 'stromatolitic' crusts. Wright feels he has recognised comparable structures in Dinantian crusts in South Wales and has accordingly suggested these as ancient examples of 'subaerial stromatolites'. His conclusions, however, are based on minimal petrographic evidence which may not warrant such precise interpretation.

Light microscope and S.E.M. studies of the Anglesey crusts proved disappointing. Few structures on the scale required were observed at all and none which were unequivocally comparable with those of Klappa and Krumbein and Giele. This may reflect the highly indurated nature of, and subsequent recrystalisation suffered by these more ancient crusts. Further inviestigation is undoubtedly required, but at the moment there is no positive evidence to support a 'stromatolitic' origin. Certainly those crusts which appear to have formed beneath a bentonitic soil cover or which line fissures in the host limestones could not have originated in this way.

In general most laminated crusts, both ancient and modern, are referred to pedogenic processes and are laminar calcretes *sensu stricto*. The main debate concerning such deposits is whether they result from in situ replacement of their host limestones or are developed by accretion on lithified bedrock either beneath a soil cover or on exposed limestone surfaces.

James (1972) favoured an in situ replacement mechanism for the formation of Recent and Pleistocene crusts from Barbados. Solution, brecciation, recrystalisation and micritisation are cited as active processes within the upper few metres of emergent Pleistocene limestones often beneath an overlying soil cover. James did not discuss the origin of internal lamination within the crusts, but Harrison (1977) working on these same deposits distinguished massive or poorly laminated subsurface micritic stringers from well laminated surface crusts. Only the former features were considered replacive in origin the latter crusts it was argued resulted from accretionary growth.

The ghosts of ooids and other allochems within laminated crusts from Florida persuaded Kahle (1977) to argue for formation by in situ micritisation of the host Pleistocene limestone. His work is important in recognising a mechanism by which such alteration could take place. He describes 'sparmicritisation' wereby low Mg calcite of either skeletal or cement origin undergoes degrading recrystallisation to micrite, a process hitherto thought unlikely in the realm of vadose diagenesis (Bathurst, 1975, p.477). Kahle argues that sparmicritisation is brought about by the action of waters enriched in chemicals released during bacterial decomposition. He further suggests after McCunn (1972), that the lamination within the crusts is caused by differential organic staining by these same solutions during the micritisation process.

Many of the textures and fabrics described by James and Kahle have been observed within the Anglesey palaeokarstic profiles both in the host limestones beneath the crusts and in the crusts themselves. Some of the more vaguely laminated examples and some of the subsurface stringers may indeed have resulted from in situ alteration (c.f. Harrison, 1977). In general, however, whilst they were undoubtedly affected by later in situ alteration processes, the finely laminated and banded nature of the Anglesey crusts is thought to reflect an essentially accretionary mode of formation, an interpretation strengthened by the occurrence of textural banding with abrupt contacts; of bands of pelloids and calcrete ooids which evidence a ditrital origin; and of internal erosional discontinuities (c.f. Multer and Hoffmeister, 1968).

The Anglesey crusts are therefore compared with laminar calcretes which accumulate on rather than within the bedrock, generally beneath a soil cover, but also on exposed surfaces (op. cit.; Read, 1974 and Harrison, 1977). The latter two authors stress the need for a lithified substrate on which the calcrete laminae can accrete. This is produced during the early stages of calcretisation by both coating of grains and micritisation i.e. by in situ alteration, reducing primary porosity until an impervious 'plugged horizon' (Gile et al, 1966) is formed. This prevents or at least retards the percolation of surface waters into the host limestones. These waters collect above the impermeable barrier and deposit the successive calcrete (micrite and/or cryptocrystalline calcite) laminae which form the crusts. The fine grain size of such calcrete deposits is thought to reflect precipitation from rapidly evaporating solutions (James, 1972). The colour and textural variations of the laminae reflect changes in organic content, rates of deposition and undoubtedly as yet poorly understood biological controls by soil microflora. Chemical considerations in calcrete deposition are summarised by Read (1976).

The broader textural variations displayed by the Anglesey crusts are closely comparable with those recognised in the Florida crusts by Multer and Hoffmeister. Crusts, or bands within crusts, composed of fine closely spaced laminae are analogous to the latter authors' dense laminated crust. The compact nature of the laminate, the lack of detrital particles and the paucity of root molds (rhizoliths) led Multer and Hoffmeister to suggest formation of such crusts beneath only a thin soil cover or indeed on exposed rock surfaces. Rhizolith-infested, fenestral and detrital bands within the Anglesey calcretes are more similar to the porous laminated crusts of Multer and Hoffmeister which they argued formed beneath thicker, tropical forest soils. Roots (rhizoliths) and soil particles (pelloids, calcrete ooids, even whisps of bentonitic clay in the case of the Anglesey examples) are incorporated within the growing crust and disrupt the finer lamination. Fenestral and brecciate fabrics represent desiccation and shrinkage effects related to frequent wetting and drying.

The sequence of textures observed within some Anglesey crusts may therefore reflect changes in the overlying bentonitic soils. Basal bands of finely laminated crust material suggest deposition on exposed bedrock surfaces following emergence and during the initial stages of soil accumulation. Overlying more 'porous' crust material indicates aggradation of the soil and its colonisation by vegetation (c.f. Multer and Hoffmeister, p.188). More varied textural sequences may result from deflation of the soil cover or may reflect more closely changes in the vegetation it supported.

(d) Rhizoliths

(i) Description

Rhizoliths are a ubiquitous feature of the Dinantian palaeokarstic horizons on Anglesey. They constitute important elements of laminated crusts, but also abound within the upper metre or so of underlying host limestones and are important indicators of emergence in the absence of surface crusts. Rhizoliths are observed, in hand specimen, as tubes, circular in cross section and rarely more than 1 mm in diameter, filled generally by sparry calcite, but more rarely by ferroan dolomite, chalcedony or sandstone (Plate 25). The latter commonly exhibits a geopetal distribution (Plate 28b). Axial fill material is invariably confined within micritic sheaths up to 2 mm wide with sharp inner contacts, but which exhibit gradational contacts with surrounding host material. The inner parts of these micritic surrounds are generally dark brown to black in colour, becoming progressively paler outwards. In exceptional examples the central channel attains a diameter of up to 3 mm and the surrounding micritic cylinder may be up to 3 cms across and distinctly banded (Plate 26a). These larger features have been observed penetrating host limestones to a depth of over a metre below palaeokarstic surfaces, whilst the more common smaller tubes can rarely be traced for more than 2 or 3 cms.

Within laminated crusts these spar filled structures display a dominantly horizontal orientation (Plate 26b) whilst in underlying host limestones repeated branching generates ramifying networks of tubes with no preferred orientation (Plate 25).

In thin section the micritic cylinders are seen to be often composite in character comprising an inner series of concentric structures and an outer micritised zone which fails distally and in which grains of the host material are present (Plate 25b). The inner features consist of vague, irregular laminae of micrite, microspar fibrous and blocky calcite (Plate 27). These line the sides of the tubes 'building' inwards towards the centre, but rarely occluding the central spar filled channel completely. These inner laminae can generally be distinguished from the surrounding micritised zone and appear to be axial features of the tubes. In other cases, particularly within laminated crusts where there is little contrast between the materials involved, the contact between axial laminae and peripheral micritised halos is difuse and illdefined (Plate 27b).

The sparry central channel of the tubes is commonly subdivided into cell-like chambers, up to 0.5 mm across, by a meshwork of micritic walls around 50 μ thick. In transverse section these walls may radiate from a central node, whilst in longitudinal sections through the tubes a simple ladder-like structure is displayed (Plate 28a) Calcite cement fills of these chambers may consist of either a single crystal or equigranular mossaic. Early fringe cements are often observed on the micritic walls.

(ii) Interpretation

Tube-like structures within calcrete profiles may be referred to the term pedotubules of soil terminology (Brewer, 1964). Such structures form in response to two main processes, burrowing by soil dwelling fauna and penetration by plant root systems. Distinction between these is not always easy, particularly in ancient sequences where a three dimentional knowledge of the structures is often lacking. Pedotubules formed by burrowing activity are commonly lined by faecal pellets (Braithwaite, 1975).
Riding and Wright (1981) and Wright (1981) have recognised burrow systems in Dinantian calcretes in South Wales using such criteria. Ovoid pelloids probably of faecal origin do occur within the Anglesey crusts, but are seldom concentrated along discrete channels. Whilst such an origin cannot be dismissed the Anglesey structures compare more closely with branching spar filled tubules, described from both Recent and other ancient calcrete profiles, thought to represent calcified root channels (e.g. James, 1972; Read, 1974; Walkden, 1974; Perkins, 1977). Little detailed analysis of such structures has been

attempted and their interpretation hitherto appears to have been largely intuitive. Recently, however, Klappa (1980b) has presented a detailed account of root-related structures from Quaternary calcretes in the western Mediterranean and this has allowed the more precise interpretation of the Anglesey examples. Klappa has suggested the term 'rhizolith' for all root related structures including root moulds, root casts, root tubules, rhizocretions and root petrifactions (op. cit. for definitions).

Rare spar-filled tubes displaying little or no alteration of the host limestone would appear to represent straight forward cement fills of vacated root voids and therefore correspond to simple root casts of Klappa. The majority of the Anglesey rhizoliths however exhibit some evidence of alteration of adjacent host limestone and/or calcification of root tissue and therefore represent rhizocretions and/or root petrifactions respectively.

The almost ubiquitous micritic halos of the Anglesey rhizoliths, which fade into surrounding host material, are evidently replacive in origin and represent rhizocretions. Klappa has described rhizocretions as " . . . pedodiagenetic accumulations of mineral matter around roots. Accumulation, usually accompanied by cementation may occur during life or after death of plant roots.". They result from centrifugal micritisation of surrounding host material in response to complex biochemically controlled solution/precipitation reactions perhaps including the sparmicritisation of Kahle (1977). Rhizocretions may also be analogous to neocalcitans and quasicalcitans of Brewer (1964). These are concentric deposits of micrite or cryptocrystalline calcite developed around voids, including root moulds, within soil profiles. They are precipitated from solutions migrating either from the host material towards the walls of voids which act as drying surfaces; or in the opposite sense by penetration of the host by solutions present within the voids. Brewer felt that the latter process would be aided by alternating wetting and drying of the sediment (p.298). Death and complete decay of the root is inferred.

Calcification around large tap roots is thought to have generated the large rhizocretion structures recorded from the Anglesey palaeokarsts e.g. Plate 26a (cf Perkins, 1977). Rhizocretions which exhibit no evidence of root petrifaction and are filled by sparry calcite, dolomite, chalcedony or indeed geopetally displosed sandstone are also in part root casts. They demonstrate the formation of the rhizocretion followed by the complete decay of the root and the filling of the resultant void by either various cements or by sand grains infiltrating from above (Plate 28b).

Concentric and chambered structures within the axial parts of rhizoliths are comparable with root petrifaction phenomena described by Klappa and also noted by Braithwaite (1975) from Holcene palaeosols on Aldabra. Klappa defines root petrifaction as ". . . a process which involves replacement, impregnation, encrustation and void filling of organic matter by mineral matter without total loss of root anatomical features". In calcretes this essentially involves the calcification of root tissues.

The concentric structures displayed by many Anglesey rhizoliths appear very similar to those illustrated by Klappa in his fig.6d. In these Recent rhizoliths such structures reflect the original concentric layering of root cells and are formed, during root decay, by the selective petrifaction of the middle lamellae which separate them. Middle lamellae of modern plant cells are rich in calcium (in the form of calcium pectate) and this is thought to promote their preferential calcification. Such preservation is the most common form of root petrifaction within Klappa's Mediterranean calcretes. The concentric layering within some of the Anglesey rhizoliths, however, is not dissimilar to that in overlying (or indeed surrounding) laminated crusts and may therefore be unrelated to root petrifaction but simply result from laminar calcrete deposition within vacated root voids. Here there are strong similarities with the calcareous cutans (calcitans) of Brewer (1964, p.216).

The meshwork of micritic walls which subdivide many Anglesey rhizoliths may represent calcified cell walls. The large size and undifferentiated nature of the 'cells' reflecting the more primitive character of some Carboniferous root vascular systems? Alternatively these structures may result from repeated partial collapse of, and micritic deposition on decaying root tissues, a mechanism favoured by Adams (1980). In either case these would represent further root petrifaction phenomena as defined by Klappa. Similar structures, however, may be formed after root decay and reflect the effects of gas bubbles within the resulting voids (c.f. Braithwaite, 1975, fig.7).

Klappa (1980b) has further proposed the term 'rhizolite' for " . . . a rock showing structural, textural and fabric details determined largely by the activity, or former activity of plant roots". On Anglesey rhizolite lithologies are represented by the rhizolith-rich surface crusts and by the highly infested, calcretised upper portions of host limestones. Perkins (1977) has described similar 'root-rocks' associated with Pleistocene palaeokarstic horizons in Florida. Rhizolite bands and indeed individual laminae within Anglesey surface crusts are distinctive for the dominantly horizontal attitude of their component rhizoliths. This is interpreted as a kind of 'plant-pot' effect caused by downward growing roots impinging on, and being deflected by indurated crust material. The spongiostrome texture of such rhizolite rich units (Plates 26b,27c) is remarkably similar to Recent root dominated peats illustrated by Cohen (1972) and suggests that some crust bands may result from the calcification of root infested peaty layers within soil profiles. Braithwaite (1975) has described similar textures from Quaternary terrestrial deposits on Aldabra.

(e) Calcrete Ooids

(i) Description

Calcrete ooids are not present within all palaeokarstic profiles, but where they do occur they commonly comprise calcrete ooid grainstone (Plate 30). These may gradationally overlie skeletal and pelloidal grainstones of underlying host material, or impersistently veneer laminated crusts and fill cracks and fissures within brecciated host limestones and indeed in earlier deposits of calcrete ooid grainstone (Plates 23b,29). Such grainstones often exhibit a degree of 'under-packing' (Plate 30) with micrite cemented grains bridging spar filled voids, the latter often floored by geopetally arranged calcrete pelloids. Detrital bands within laminated crusts may be largely composed of calcrete ooids (Plate 20).

Calcrete ooids up to 2 mm in diameter consist of a nucleus derived from the host limestones (skeletal grain, pelloid, or intraclast) surrounded by an often uneven coating, up to 1 mm thick, of dark brown to cream micrite, cryptocrystalline calcite, and microspar (Plates 29,30). Coatings are generally structureless but also exhibit pelloidal or clotted textures and may display a vague concentric lamination. They often contain fine skeletal debris and even quartz grains indicating the incorporation of larger detrital particles. Where the nucleii are elongated the micritic coatings are thickest normal to the long axis of the grain (Plate 30) a feature also noted by James (1972). Compound calcrete ooids are common. Larger calcrete pisoids i.e. diameter > 2 mm are developed generally because of the large size of the nucleus, rather than the thickness of the calcrete coating. Pisoids with thick well laminated coats as described by Read (1974) have not been recorded from the Anglesey palaeokarsts.

(ii) Interpretation

The similarity of the micritic coats to the micritic laminae within laminated crusts and the incorporation of the ooids within the crusts demonstrate their close affinity. The ooids are developed by calcrete deposition around individual grains rather than as laminar sheets as in the crusts. They are identical to the calcrete ooids of Read (1974) who argues that such particles form by micrite deposition around loose uncemented grains derived from underlying host limestones and incorporated within overlying soils. Coalescence of the ooids is prevented by the gradual movement of the grains due to either soil creep, or the activity of roots or soil dwelling faunas. This allows calcrete to be deposited on all sides and leads to the formation of concentric coats.

Read also noted, however, that very similar features were developed by calcrete deposition around grains within the host limestone in his 'mottled zone'. These intra-host calcrete ooids are closely comparable with the coated particles of James (1972), the diagenetic ooids of Siesser (1973) and the laminated grains of Harrison (1977). It is difficult to see, however, with the original grains in close contact, how discrete calcrete coats are precipitated around individual nucleii. Part of this space problem may be resolved if the micritic coats are partly replacive in origin as in Read's 'mottled zone' and in some of James' examples. Whilst Seisser argues that calcrete precipitation may physically push grains apart, essentially a force of crystallization effect. Part of the problem may rest with the distinction between soil and host. At what depth does an unconsolidated carbonate sediment, possibly colonised at the surface by plants cease to be a soil? Clearly there will be a critical depth of overburden beneath which constituent grains will be essentially immobile. These lower particles are perhaps unlikely to provide nucleii for accretionary calcrete ooids though possibly for ones originating from replacement. The effects of early lithification are obviously critical in this context also.

Within the deeper parts of Dinantian palaeokarstic profiles calcrete ooids are rare, micrite occurs as a centrapetal replacement or interstitial cement (Section 3.4h) rather than as individual coats around discrete grains. Well developed calcrete ooids, comprising grainstone units, occur only in the upper parts of palaeokarstic profiles i.e. where the allochem nucleii where potentially mobile, and therefore appear to have formed in accord with Read's initial ideas. Cessation of movement allowed micrite deposition between grains (cementation) first leading to the formation of compound ooids and eventually to their inclusion within laminated crusts. Transitional stages in this process involved the formation of micritic bridges between grains (c.f. intertextic fabric of Brewer, 1964). The underpacked texture often displayed by calcrete ooid grainstones is thoughtto reflect the continued settling or creep of the grains during this stage.

(f) Brecciation textures

(i) Description

The effects of repeated brecciation are evident throughout palaeokarstic profiles occurring within host limestones, laminated crusts and blocks and nodules within the palaeosols. Combined with the effects of concomitant cementation and recrystallization, polyphase brecciation has led to the complex textures illustrated in Plates 31, 32, 33. Many of these compare closely with those described by Freytet (1973) from Cretaceous and Eocene limestones in France and which are similarly associated with penecontemporaneous emergent episodes.

The Anglesey textures are largely the result of in situ brecciation with only limited reworking of the clasts within the basal regolith of the palaeosols. In hand specimen the effects of brecciation range from simple horizontal or vertical sheet-like fissures to complex exploded jigsaw-like patterns (Plate 33a). The breccia clasts are generally angular, but subsequent dissolution and particularly later pressure solution may cause rounding and impart a more rubbly appearance to the rock (Plate 31c). The edges of the clasts vary from sharp to diffuse, the latter due to subsequent recrystallization, indeed care has to be taken to distinguish true breccias from strikingly similar textures produced by patchy recrystallization (pseudobreccias). Cracks and fissures may be filled by coarse sparry calcite cements, 'chalky' micro- or pseudospar or by detrital particles, including calcrete ooids or sand grains. Rhizoliths have been observed within fill material (Plate 31b). Caliche-type nodules of microspar commonly display radial and concentric cracking of a septarian type (Plate 32a) or exhibit distinctive stellate, sigmoidal or simple lozenge shaped gashes at their centres (Plates 32b,c; 34b).

In thin section several smaller scale brecciation phenomena are apparent. The retreat fissures and bursting effects of Freytet (1973) are well seen and comprise curvilinear fissures commonly developed around allochems (skeletal grains, ooids etc.) within calcretised limestones and which separate these particles from their surrounding matrix material (cf. James, 1972; Harrison and Steinen 1978). Similar fracturing also occurs around and across 'clasts' of unaltered limestone within a recrystallised matrix (Plate 33a,b) whilst ghosts of such fractural 'clasts' demonstrate the close association of brecciation and recrystallization.

Fenestral structures within laminated crusts may also be regarded as a brecciation phenomena and comprise irregular tapering fractures orientated sub-parallel to the lamination (Plate 24b). Fills are generally of coarse sparry calcite although geopetally arranged internal sediment is not uncommon.

Quartz pebbles caught up within palaeokarstic profiles often exhibit cracking and spalling effects, with the resultant fissures filled by coarse sparry calcite cement (Plate 33c).

(ii) <u>Interpretation</u>

The incorporation of breccia clasts into palaeosol regolith and the filling of fissures by calcrete ooids or sand grains demonstrate the contemporary nature of the brecciation. Care must be taken however to distinguish calcite veining of tectonic affinity which is evident throughout the Dinantian sequence. The gross cross cutting relationships of these later veins is generally apparent.

Most synsedimentary brecciation effects within calcrete profiles are commonly explained in terms of expansion and subsequent shrinkage related to wetting and drying of the sediment. Sheet-like cracks and retreat fissures around nodules and allochems may well have developed in this way and are analogous to the skew and craze planes of Brewer (1964). Fenestral structures within laminated crusts are identical to 'birdseye' structures developed in supratidal algal mats and carbonate muds (Shinn, 1968) where they represent desiccation and shrinkage phenomena. Whilst the environmental setting is somewhat different it seems likely that the fenestrae within the crusts are formed by these same processes thus bringing into doubt the value of birdseye structures as environmental indicators.

Septarian type cracking within nodules, of which the stellate and lozenge shaped fractures appear insipient forms, are thought to represent further desiccation phenomena (Pettijohn, 1957), but their exact origin is poorly understood and may involve diagenetic effects related to osmosis (Brewer, p.279) or syneresis.

Brecciation and the formation of tepee structures within calcrete hardpans has been described by Assereto and Kendall (1977) and was thought to result from the force of crystallization of growing calcite cements. Cementation and recrystallization were clearly active within Anglesey palaeokarstic horizons during brecciation and force of crystallization may therefore have been a contributory factor in the latter's development. Displacive calcite spherulites have been observed within the Anglesey calcretes (Section 3.4g) and demonstrate that such forces were active.

The spalling of quartz pebbles within palaeokarstic profiles probably takes place along microfractures within the grains developed during their transportation. These planes of weakness may be exploited by exfoliation type processes (wetting and drying or heating and cooling) and/or by the force of crystallization of calcite cements precipitated within the cracks.

Klappa (1980a) has emphasized plant roots as agents of brecciation within Quaternary calcretes. Certainly roots were active within the Anglesey breccias as evidence by the presence of rhizoliths (Plate 31b). These would undoubtedly have aided in the brecciation process by penetrating along pre-existing cracks and fissures, but to what extent roots were responsible for the initial brecciation as Klappa proposes is difficult to say.

(g) Recrystallisation effects

The effects of penecontemporaneous recrystallisation within the palaeokarstic profiles are complex and diverse and cannot be detailed in full in the space available here. Principal effects include the growth of clots and nodules of micro- and pseudospar; and of radial fibrous calcite both as coarse, bladed fringes to the above clots and nodules but also as distinctive needle crystal sperulites.

(i) Growths of micro- and pseudospar : description

These growths comprise clots of microspar and occasionally of coarser pseudospar. Individual clots are around 3 to 4 mm across, but coalesc to form cauliflower-like tufts (Plate 34a) which may in turn combine to give larger nodules with a distinctive nobbly form (Plate 34b). The clots within these tuffs and nodules are often highly contorted folding over on themselves in a manner reminiscent of enterolithic structure in evaporites (e.g. Shearman and Fuller, 1969). In the field these various growths exhibit a characteristic white weathered 'chalky' appearance and texture.

In many places the clots and tufts exhibit fringing growths of fibrous calcite (Plate 35) with microspar centres passing into pseudospar and then with gradual elongation of the crystals into the often coarse bladed calcite crystals which comprise the fringes. The latter are locally up to 5 cm long. These structures are closely comparable with the stellate masses of radial fibrous calcite described by Orme and Brown (1963) and Bathurst (1975, fig.332). The fibrous fringes may grow from all sides of the microspar clots but appear often to be preferentially developed and thickest on downward facing parts.

Centres of microspar clots and tufts commonly display sigmoidal and lozenge shaped spar filled gashes (Plate 34b; see (f) above). Clots may envelope and coat skeletal grains, but in other cases ghost of such grains evidence replacement.

These various phenomena are commonly developed within rubbly palaeokarstic profiles where they display intimate relationships with other alteration effects. 'Chalky' nodules of microspar are often brecciated, whilst microspar also heals cracks and fissures of altered limestone and cements rubbly blocks of laminated crust (Plate 31c). Clots and nodules of microspar are not however confined to strata immediately below palaeokarstic surfaces but occur pervasively throughout minor cycles, invading host limestones and obliterating primary textures. They are particularly prevalent within strata with intercalated shale bands e.g. Flagstaff Formation (Plate 76) in which thin beds of limestone may be totally composed of such clotted material. Within the shales lamination is distorted around the nodules and combined with the non-crushed nature of contained skeletal grains demonstrate an origin prior to compaction. Microspar growths are also common within palaeosol material which has been reworked during marine transgression and is mixed with overlying limestone strata i.e. at the base of minor cycles. Here clots and nodules of microspar envelope fossils or selectively grow along and pick out burrow forms.

(ii) Interpretation

The widely developed nature of these various phenomena ranging from the basal layers of minor cycles as well as palaeokarstic profiles, as nodules within shales and as replacive growths within grainstones (see Section 5.5c) remains one of the perplexing features of the Dinantian sequence on Anglesey.

For the chalky nodules of micro- and pseudospar occurring within rubbly palaeokarstic profiles the intimate relationships with other alteration phenomena e.g. brecciation, demonstrate a contemporary origin. They appear identical both in form and composition and in geological setting with now widely documented caliche nodules (e.g. Reeves, 1970), diagenetic accumulations of $CaCO_3$ developed within soils. They resemble closely the cornstones within the Anglesey Old Red Sandstone also assigned a pedogenic origin (Allen, 1965).

Despite the many obvious similarities the origin of the clots and nodules of microspar developed within the lower parts of minor cycles is less certain. They cannot easily be related to contemporary emergent effects and their interpretation rest on more circumstantial evidence. During the tenure of this research several origins for these phenomena have been considered.

Where the clotted structures are exposed coating bedding planes e.g. Plate 34a, they bear a striking resemblance to modern tufa deposits observed forming along parts of the Anglesey coast. Closer comparison still is possible with the coniatolites of Purser and Loreau (1973) described from the Persian Gulf. These comprise aragonitic encrustations on supratidal beach rocks, but also coat shells and line cavities and produce surface textures (cf. op. cit. fig.19) similar to the microspar growths in the Anglesey Dinantian. Internally however there are obvious differences and the Anglesey features fail to display the concentric banding and totally radial fibrous composition of both modern tufa and coniatolites. But could the Dinantian structures have resulted from calcitisation of such aragonitic growths? Neomorphic inversion of skeletal aragonite to calcite tends to give often quite coarse sparry mosaics (Bathurst, 1975 p.486) and appears not to result in the extensive development of microspar. The same effects also characterise the inversion of inorganic aragonite e.g. modern ooids (Sandberg, 1975; see Section 5.5d).

The microspar clots were also compared with algal thrombolites (Aitken, 1967). These form biohermal and biostromal deposits and are formed in response to the trapping and binding activities of blue-green algae, but lacking the distinctive cryptalgal lamination of stromatolites. Aitken describes thrombolites as "characterised by a macroscopic clotted or spongy fabric with external surfaces pimpled, corrugated or pitted" whilst the microfabric "consists of centimetre-sized patches or clots of microcrystalline limestone". Logan et al (1974) also refer to thrombolitic textures produced by Recent algal mats in Shark Bay. These too appear similar to the Anglesey clotted limestones especially in their potential to generate extensive sheets of such deposits comparable with those observed in the Flagstaff Formation.

Despite these similarities there are again serious problems with such an interpretation. The modern thrombolites of Logan et al exhibit well developed fenestral fabrics absent in the Anglesey clots and nodules. Whilst in both these Recent and ancient occurrences, thrombolites form only one of a suiteof cryptalgal structures and are intimately associated with more readily recognised stromatolitic forms. The latter structures are not recorded from the Dinantian clotted limestones. Of greater significance however is the occurrence of microspar clots and nodules completely enclosed in shale or replacing burrows and on the undersides of fossils, habitats incompatible with the growth of photosynthesizing blue-green algae.

The similarity of the contortions within the cauliflower-like tufts of clotted material to enterolithic structure in anhydrite prompted investigation into a further possible origin, from evaporite replacement. Certainly as well as the gross morphological similarities, the occurrence of the clots and nodules as precompactional growths in shales and replacing primary carbonate textures is compatible with an evaporitic origin (Shearman and Fuller, 1969). The occurrence of collapse structures (see Section 3.5a) and of length slow chalcedony (Folk and Pitman, 1972) within the succession is also in keeping with the former presence of evaporites, whilst more direct evidence is provided by the recognition in some limestones of gypsum pseudomorphs (Section 5.5e). Again however problems exist. The microspar of the Dinantian clots differs from the coarse sparry mosaics which normally result from the calcitisation of anhydrite (Shearman and Fuller, 1969), whilst the presence of skeletal grains is also at odds with an evaporitic origin, such material is usually completely dissolved during the initial phase of evaporite growth (op. cit.).

The evidence both positive and negative suggests that the clots and nodules of microspar with their locally developed fibrous fringes are diagenetic growths, replacive and perhaps within shales also displacive in origin. Moreover they appear to have achieved their present form as calcitic growths and not to have replaced an earlier mineral species or polymorph. In shales such structures grew prior to compaction and formation during early, shallow burial diagenesis is demonstrated. For replacive growths of microspar within intercalated limestones e.g. Flagstaff Formation, an early origin must also be suspected, but in other units the distinction from deep burial neomorphic effects will perhaps only be achieved with detailed isotopic 'finger-printing'.

The occurrence of nodules of clotted microspar within reworked palaeosol material is significant at this point. They appear very similar to the caliche nodules within rubbly, in some places underlying palaeokarstic profiles, yet they envelope skeletal material which is part of the next marine phase of deposition and are therefore unrelated to underlying palaeokarstic horizons. If indeed they do represent some sort of caliche growth they could only have grown during later periods of emergence i.e. related to the next higher or indeed subsequent palaeokarstic surfaces.

The evidence is suggestive rather than conclusive, based on 'look-alike' criteria rather than firm factual data, but indicates that not only the pervasive growths of microspar but perhaps many of the Diagenetic transformations within the Dinantian limestones of North Wales (Bathurst, 1958, 1959 and Orme and Brown, 1963) may be very early. Possibly related to the frequent emergence suffered by these minor cyclic sequences. The thicker developments of fibrous fringes on the lower sides of microspar clots may indicate an origin for these features related to downward perculating vadose groundwaters. The prevalence of clotted growths within units with intercalated shales e.g. Flagstaff Formation,and in reworked palaeosols may suggest that their formation was favoured by such vadose solutions meeting impermeable layers and that these diagenetic effects are perhaps related to perched water tables and/or zones of phreatic/vadose transition.

(iii) Needle crystal spherulites : Description

Within several rubbly palaeokarstic profiles, most notably that at Pedolau at the top of the minor cycle of the same name radial fibrous calcite occurs as distinctive needle crystal spherulites (Plates 36, 37) which locally join and develop fibrous sheets. The spherulites, up to 1 cm in diameter, may be isolated within a matrix of haematitic silty, sandy, even conglomeratic mudstone. Elsewhere they have grown so close together to have interfered with one another's growth and have developed fitted polygonal outlines, excluding matrix material completely (Plate 36). In these latter cases hand specimens exhibit a pelletal or botryoidal texture depending on the size of the structures.

The spherulties are composed of fine, radiating, inclusion rich needles of calcite (Plate 37c). Outer portions may exhibit secondary tufted growths (Plate 36b) or may be fringed by more blocky calcite crstals (Plate 37b). In other cases the edges of the spherulites are ill-defined with isolated needles penetrating and fading into the surrounding matrix (Plate 36c). Centres consist of more equant crystals, but no particulate nucleii have been observed. Around the centres the constituent calcite needles have generally become fused together to form single crystals which exhibit curved cleavage planes (Plate 36c) and undulose extinction. These portions of the spherulites appear identical to the fascicular-optic calcite of Kendall (1977). The spherulites as a whole display characteristic uniaxial-cross extinction patterns (Plate 37a).

The spherulites often contain scattered quartz grains identical to those occurring in sounding matrix material. In many instances however quartz grains are also concentrated around the outer edges of the spherulites (Plate 36b). The effects of penecontemporaneous brecciation may cut across the spherulites randomly but often takes the form of concentric shrinkage cracks (Plate 37c).

Staining of the calcite needles which make up the spherulites by acidic solutions of potassium ferricyanide and Alizarin Red S gave shades of purple. This contrasts with the royal blues of later void filling and vein calcite, but both responses are indicative of ferroan calcite (Dickson, 1976).

(iv) Interpretation

Peripheral needles penetrating surrounding matrix material on the one hand, and the mutual interference of spherulites to give fitted, polygonal outlines on the other both evidence an origin from in situ growth rather than as detrital particles. Nor do the Anglesey spherulites exhibit the well developed concentric structure of Dunham's (1969) vadose pisolites. Spherulitic structures which have originated in situ are not widely described from calcrete deposits. Void filling growths of radial fibrous calcite have been described from Quaternary calcretes by Watts (1980), whilst the cone-in-cone sheets are reminiscent of the crystal sheets of Brewer (1964). Minute spherulitic structures related to the calcification of lichen and plant tissues have been described by Klappa (1978, 1979) but are clearly unrelated to the larger Anglesey features. Adams and Cossey (1982) briefly mention spherulites with extinction crosses from Dinantian calcretes in Northern England. Spherulitic and other radial fibrous growths are more commonly reported from speleothem deposits where they are invariably interpreted as void filling cements of one sort or another (e.g. Assereto and Kendall, 1977; Assereto and Folk, 1980; Chafetz and Butler, 1980). The Anglesey examples may have been initiated as void filling growths but the inclusion rich nature of the spherulites suggest a replacive origin whilst the occurrence of fine needles penetrating surrounding matrix and the concentration of matrix quartz grains around the edges of some spherulites argue strongly for a displacive origin. Displacive growths of radial fibrous calcite have been described by Folk (1965).

Such displacive growth mechanisms are favoured by, if not dependent on the unconsolidated nature of the surrounding matrix, and are likely to be of early diagenetic origin. The ferroan composition of the calcite in the Anlgesey spherulites may, at first, seem at odds with this conclusion (Bathurst, 1975 p.432), but given the ferruginous nature of the matrix material a ferroan composition is hardly surprising. Moreover the purple stained ferroan calcite of the spherulites is readily distinguished from the blue stained later void filling and vein ferroan calcites.

The acicular morphology of the crystals within the spherulites may reflect the chemical environment in which they were precipitated but it has also been suggested that such habits may be inherited from an earlier mineral phase. Kendall and Broughton (1977) have sensibly pointed out that "any crystal, of whatever composition, if allowed to grow continuously between similarly orientated neighbours will assume a fibrous habit". Yet the distal crystals of the spherulites lack such confinement and demonstrate the acicular habit to have been a primary one.

Folk (1974) has argued that such needle habits in calcite are due to 'Mg-poisoning' during crystal growth and reflect precipitation, from solutions with elevated Mg/Ca ratios. Needle crystals of calcite in Barbados caliche (James, 1972 and Harrison, 1977) are precipitated from meteoric solutions enriched in Mg^{2+} ions, the latter leached from the base rich montmorillonitic soils with obvious implications for the bentonitic palaeosols on Anglesey.

Watts (1980) has suggested that the aciciular morphology of crystals of low-Mg calcite within Quaternary calcretes indicates an original high-Mg calcite mineralogy crystals of which often display such habits. Kendall (1977) felt that fascicular-optic calcite such as characterises the inner portions of the Anglesey spherulites, may form simply by the coalescence of adjacent needles of low-Mg calcite. He further suggested, however, that it may result from the inversion of acicular bundles of aragonite (by calcitisation) or high-Mg calcite (by loss of magnesium). There is no evidence e.g. square ended crystals or equant mozaic inversion textures (Folk and Assereto, 1976; Assereto and Folk, 1980), from the Anglesey calcretes to suggest the former presence of aragonite, but former high-Mg calcite presursors cannot be ruled out.

The "Mg-poisoning' of Folk, whilst influencing crystal morphology, does not necessarily result in a high-Mg mineralogy. Folk and Assereto (1976) record acicular calcite crystals with Mg contents well below the range of high-Mg calcites (Bathurst, 1975). James has also suggested that high concentrations of various other ions within the precipitating medium, notably SO_4^{2-}, may also promote acicular calcite habits in caliche. With reference to the Anglesey spherulites, therefore, it seems likely that the complex chemistry of solutions within the zone of pedogenic alteration was a major factor in influencing crystal morphology. Whether these abnormal crystal habits reflect a previous high magnesium mineralogy as Watts suggest remains uncertain.

(h) Early Cementation

The circumstantial evidence that strata beneath palaeokarstic surfaces suffered early cementation is unequivocal and includes the steep, locally overhanging sides developed by surface hummocks and pipes, penecontemporaneous brecciation, the abundance of lithoclasts both of host limestone as nucleii to calcrete ooids and of altered limestone within basal beds of overlying strata and the occurrence of Trypanites borings (Section 3.5b).

Micrite precipitated as cement between grains rather than as a replacement is a ubiquitous feature of calcretised limestones and its textures and fabrics provide the clearest petrographic evidence for early cementation. The results of micritic cementation range throughout palaeokarstic profiles from deep within host limestones to surface crusts and calcrete ooids. Concentric micrite coats around the latter particles may be compared with the aragonitic cortex of modern marine ooids and thought of as cements frustrated by the movement of the grains (Bathurst, 1968). Compound calcrete ooids and the incorporation of calcrete ooids within laminated crusts demonstrate the eventual success of micrite cementation over these grains. More obvious cement textures are observed within host limestones, particularly grainstone in which micritic cements are readily identified. In packstone lithologies such fine grained cements may easily be confused with primary detrital micritic material; indeed micrite cements (often in parts recrystallized to microspar) may envelope grains and occlude original pore space completely giving secondary packstone or, where replacement micritisation has also been active, wackestone textures (c.f. Read, 1974). In other cases micrite cements are localised at grain contacts and display typical vadose meniscus structures (Plate 30; c.f. 'hour-glass' structures of Perkins, 1977). Micrite cement also fills skeletal cavities such as chambers within foraminifera, but its deposition invariably predated molluscan aragonite dissolution since the moulds of such grains are now filled by sparry, ferroan calcite. Lithoclasts of micrite cemented host limestone often form the nucleii for micrite coated calcrete ooids and attest to both its early and continued precipitation within palaeokarstic profiles. Pore spaces which remained after micrite cementation are now filled by blocky ferroan calcite cements.

The culmination of micrite cementation within calcrete profiles was as discrete sheets on the surface of the host limestones building to form laminated crusts. Detrital grains incorporated within the crusts were coated by micrite just as vacated root voids (rhizoliths) were lined by it. Multer and Hoffmeister (1968) have suggested that micritic laminae within Recent Florida crusts may also include 'minute soil particles' washed down and incorporated within the growing crust. It is possible, therefore, that the micritic 'cements' within the Anglesey calcretes are not the result purely of micrite precipitation as cement but may also be partly micro-detrital in origin. In either case the accretionary distribution of this micritic material is closely analogous to that of calcareous cutans (calcitans) within modern soils (Brewer, 1964). These latter features may form bridges between grains (intertextic fabric) comparable to the meniscus type cement structures observed in the Anglesey profiles.

The above criteria for penecontemporaneous lithification (e.g. piping, fissuring, etc.), however, are often apparent in host limestones where interstitial micrite cements are absent. In these, and in unaltered limestones below calcrete profiles but which were still, one would suppose, within the zone of vadose diagenesis, the evidence for early cementation is often disappointing. Non-ferroan fringe cements which predate skeletal aragonite dissolution may represent early vadose cements (Plate 39) but fail to exhibit any of the diagnostic meniscus or stalactitic effects. Casts of molluscan grains and remaining parts of original pore space are filled by subsequent ferroan cements of phreatic affinity (Bathurst, 1975, p.432). Even the non-ferroan fringes are not universally developed, and original pore space may be filled totally by blocky ferroan generations.

The failure of petrography to elucidate the early cementation history of these rocks is perhaps best displayed with reference to the Trwyn Dwlban sandstone pipe locality (Section 4.2). Here there is evidence of two generations of pipes, a later sandstone filled set having in places cut and eroded an earlier set plugged by conglomerate (Plate 14). The evidence for early lithification firstly of the host limestones and secondly of the conglomeratic fills is clear. Yet in thin sections of the limestones, where they have resisted later dolomitisation, blocky ferroan cements predominate. Even cathodoluminescence has failed to distinguish early cements which may correspond to these lithification events (Walkden and Davies, in prep.). Similarly in the conglomeratic sandstones evidence of early presumably carbonate cements is missing and the quartz grains are now bound together by syntaxial overgrowths of silica.

Such disappointing petrographic results may indicate a patchy distribution of early cements which often go unobserved, or alternatively that initial cementation was achieved by cryptocrystalline cement films, or that the evidence for early cement generations has been lost during subsequent diagenesis.

3.5 OTHER MISCELLANEOUS FEATURES OF PALAEOKARSTIC HORIZONS

(a) Collapse Structures

Collapse structures have been observed at three localities, Lligwy Bay, Pedolau and Fedw-fawr. The former is the most famous and represents the Lligwy Bay Disturbance of Greenly (1919, p.615), a belt of disturbed strata up to 30 m across within the southern cliffs of Lligwy Bay [4996 8710]. This has been described more recently by Challinor and Bates (1973) and Bates and Davies (1980). The structure is illustrated in Plates 40 and 41. At its western end steeply dipping limestones apparently rest on horizontally bedded calcareous sandstones and shales (Plate 40b). Towards the centre of the disturbance thinly bedded rubbled limestones are disposed vertically (Plate 40c) and are followed towards the seaward end of structure by a zone of chaotic limestone blocks set in a matrix of red silty shales with bands of fine yellow sandstone (Plate 41a,b).The contacts both east and west with adjacent undisturbed strata are not exposed, but the cliffs to the east, seaward of the disturbance, are developed in gently dipping, unaffected rocks of the Lligwy Beds. The succession within the disturbed belt matches that of strata higher in the sequence (Porth Forllwyd Beds to Pedolau Beds) and its origin through collapse, therefore, appears certain.

The remaining, much smaller collapse structures are developed within the Pedolau Beds at Pedolau [5080 8710] and in minor cycle F6 (≡ Pedolau Beds) at Fedw-fawr [6038 8204] in Penmon (Plate 41c). Clearly there appears to be a link between all these features and the Pedolau Beds. It is tempting to relate them to the palaeokarstic phase at the top of this minor cycle; to the collapse of vadose cavern systems developed, possibly during previous palaeokarstic episodes, within the underlying limestone strata. Supportive evidence for such penecontemporaneous foundering is wanting, however, and until forthcoming, possibly through detailed petrological considerations, a much later origin (e.g. Triassic or Tertiary) cannot be ruled out.

(b) Borings

Distinctive slot shaped borings of *Trypanites* have been recorded from two palaeokarst localities, at Huslan [5220 8315] at the top of the Lower Dinas Beds and at Trwyn Du [6335 8176] at the top of minor cycle TB6. In both cases these inchnofossils are located on upstanding hummocks and the palaeokarstic surfaces are overlain by marine shales rather than bentonitic palaeosols (Plates 42 and 43). In the former locality the borings have been filled by skeletal debris which has often been selectively dolomitised and which weathers proud of the surrounding limestone (Plate 42b). The shale fills in the latter locality have often been removed by recent erosion and the now empty bores must be closely comparable to their original form (Plate 43).

The occurrence of such features on palaeokarstic surfaces clearly demonstrates the early lithified nature of the sediment, but in other ways seems at odds with the emergent setting envisaged for these horizons. Unless they represent the effects of some hitherto unreported Dinantian terrestrial boring fauna, *Trypanites* is generally regarded as a marine inchnogenus (Bromley, 1972). The well preserved nature of the Anglesey examples and their occurrence only on upstanding hummocks demonstrates formation after palaeokarstification, whilst their fill materials link them with overlying marine strata. They are thought to have been developed after transgression of the palaeokarstic surfaces during which any overlying palaeosol was removed. The exposed prominences of the surfaces, initially free of sediment and exposed on the sea bed, would then have acted as hardgrounds available for colonisation by both boring and encrusting organisms. Cerioid corals and *Chaetetes* colonies observed resting directly on palaeokarstic surfaces may fall into the latter category.

CHAPTER FOUR

SANDSTONES

.1 INTRODUCTION

Although the Anglesey Lower Carboniferous succession is a dominantly limestone sequence it is punctuated by several siliciclastic units (Fig. 7). Volumetrically these occupy a small fraction of the sequence, but excellent coastal exposures, their past neglect and the recognition of their importance in any overall stratigraphic model prompted a detailed study.

Previous work on the sandstones avoided detailed environmental analysis. Greenly (1919) confines his comments to the various stratigraphic relationships he encountered in mapping the sandstone bodies. Mitchell (1964) attempted to demonstrate possible fault control of sandstone deposition, but whilst he recognised and described some of the lithologies which make up the thicker sandstone units he neglected to give any environmental interpretation. George (1974) describes the main sandstones as "delta spreads thick and massive enough to exclude limestones altogether", an interpretation which appears more intuitive than based on a detailed examination of the sections.

The present investigation has taken advantage of the great volume of recent literature detailing the deposits and processes of modern siliciclastic environments allowing their more precise comparison with ancient sequences. Combined with detailed studies of the superb coastal sections this has resulted in the recognition, description and interpretation of recurrent lithofacies, and allowed environmental reconstruction for the main sandstone units.

.2 DISTRIBUTION

Sandstone units occur in all three areas. In the Penmon Area both the Fedw and the Parc Sandstones are poorly exposed, the latter being also highly faulted. The Fanogle, Carnedd, Edwen and Moel-y-don Sandstones of the Straits Area show several features of interest but again their limited exposure precludes a lengthy discussion. It is the sandstones of the Principal Area and the excellent sections on the NE coast which provide the main basis for this account.

.3 SETTING

Sandstones occur in two main settings; (a) at the unconformable contact between the Lower Carboniferous strata and the adjacent older rocks as a series of basal sandstone bodies; and (b) associated with palaeokarstic surfaces within the limestone succession i.e. at cycle boundaries. These latter sandstone units are distributed throughout the succession and occur in all the formations. They are particularly prevalent however in strata of Brigantian age, the Traeth Bychan Formation and the Red Wharf Formation (Fig.7) in which sandstones have been recognised at nearly every cycle boundary. Most such sandstones occur within and are largely confined to channels incised through palaeokarstic surfaces, but less common sheet sandstones also occur.

.4 LITHOFACIES

These various sandstone units are constructed, in the main, from three recurrent lithofacies:-

Lithofacies A	Quartz pebble conglomerates and coarse pebbly sandstones
Lithofacies B	Interbedded shales and thin sharp based, fine to medium grained sandstones
Lithofacies C	Calcareous sandstones

(a) Lithofacies A

(i) Description

Quartz pebble conglomerates form the distinctive lithology of this lithofacies (Plate 44). These are the "Normal Conglomerates" of Greenly (1919, p.602) for which he gives a detailed account of the derivation of the constituent pebbles. The vast majority he recognised as from the Mona Complex with up to 90% attributable to the creamy white vein and augen quartz from the Penmynydd Zone and the Gwna Green Schist. Scattered pebbles of distinctive red Gwna jasper are characteristic and other resistant rock types from the Mona Complex and also of Ordovician age are represented.

The conglomerates are poorly sorted with grain sizes ranging from medium sand to pebbles with long axes up to 13 cms. The tendency is towards a clast supported texture although this is almost meaningless in such poorly sorted deposits.

In thin section the finer fraction is observed to consist largely of quartz grains with minor biotite flakes and interstitial iron oxides and kaolinite. Grain contacts are often highly sutured due to pressure solution, but syntaxial overgrowths are also present.

Despite sutured contacts the original character of the grains is often evident. The finer quartz sands are angular, but larger pebbles generally show rounding and many, especially the augen quartzes, are well rounded. Sphericity is generally poor with bladed and roller shapes (Zingg, 1935) most common reflecting the metamorphic origins of many of the grains.

As conglomeratic units are traced laterally they often pass, quite rapidly, into areas where pebble content is reduced. Here conglomerates grade into coarse pebbly sandstones or even, with improved sorting, medium grained sandstones, though seldom without a few scattered pebbles. Contacts between these lithologies are gradational with pebble rich areas ill-defined and pebbles poorly segregated (c.f. Clifton, 1973).

Bedding in the conglomerates and pebbly sandstones is poorly defined. A crude horizontal stratification is often displayed but more commonly they appear massive and form units several metres thick in which widely spaced (1 to 2 m) impersistent silty laminae provide the only trace of bedding. The presence of siltstone clasts within the conglomerates indicates erosion and removal of some such laminae. Elsewhere discrete beds of the above pebbly sandstones up to 1.5 m thick interrupt conglomeratic sequences (Plate 44b).

Sedimentary structures observed in the conglomerates include both tabular and trough cross bedding (Plate 45a) with set thicknesses of up to 2 m recorded. These display occasional reactivation surfaces of the type described by McCabe and Jones (1977). A single example of possible epsilon cross bedding (Plate 56). has also been observed. In the finer sandstones, small scale cross lamination occurs and in better bedded areas upper bedding surfaces display poorly formed current ripples. Discrete scours up to 2 m wide and 35 cms deep also occur and may have either a pebbly sandstone or conglomeratic fill (Plate 58c).

Pebble imbrication in massive conglomerates is rarely observed directly, but measured fabrics appear to fall into the a (t) b (i) category of Harms et al (1975) with pebble long axes orientated perpendicular to the direction of palaeoflow as indicated by associated cross bedded units.

Occasional trace fossils occur in the pebbly sandstones, generally simple tracks and pit-like burrows, which are referred to the Scoyenia trace fossil assemblage of Seilacher (1967). Plant fragments are common particularly in the basal sandstones (Plate 44a) and rootlets have also been observed in the finer lithologies.

(ii) Interpretation

The predominantly coarse nature of the lithofacies and the presence of tractional sedimentary structures indicate deposition under high energy current conditions. Whilst intercalated silty laminae and finer sandstone beds demonstrate that these conditions did not prevail constantly. Tabular cross sets are the product of the migration of straight crested bed forms; trough cross sets the product of sinuous crested, liguoid or lunate forms (Allen, 1963). Such dune forms are characteristic of the upper part of the lower flow regime (Simons et al, 1965), but the lack of detailed flume work on gravel grade sediment prevents a more detailed appraisal of flow parameters.

The origin of horizontal stratification in conglomerates has been discussed by Harms et al (1975). Current velocities capable of rolling pebbles and cobbles (see below) are great enough to place sand grade sediment into suspension. Fluctuations in current strength allow these suspended fractions to settle and infiltrate the bed load producing an alternation of pebbly and more sandy layers.

The more massive portions of the lithofacies are less easily interpreted hydrodynamically. Again the coarseness of the sediment and the presence of eroded siltstone clasts suggests strong current action. The prevalence of a(t) b (i) imbrication fabrics indicates emplacement of the clasts by rolling mechanisms and implies not unexpectedly that transportation of such coarse sediment took place only within the traction carpet (Harms et al, 1975). Experimental critical velocities for the movement of some of the larger cobble grade clasts (up to 10 cms diameter) are in excess of 2 m/sec (Sundborg, 1956). The absence of internal stratification and the poorly sorted nature of the deposit probably reflect high sediment concentrations and the lack of protracted reworking processes.

Smaller scale cross lamination observed in the finer sandstones is the product of migrating ripples (Allen, 1963) and reflects deposition by unidirectional currents operating in the lower part of the lower flow regime (Simons et al, 1965). Such structures are necessarily precluded from the coarser lithologies since ripples cannot form in sediment with a mean size greater than 1 mm (Harms et al, 1975).

From a more general environmental interpretation the presence of rootlets, the abundance of plant debris and the sparse Scoyenia facies inchnofauna indicate a terrestrial setting for the lithofacies.

(b) Lithofacies B

(i) Description

This lithofacies consists of silty shales with thin sharp based sandstone beds (Plate 45b).

The sandstones are generally quartz rich with minor amounts of biotite and muscovite often concentrated along micaceous laminae. Less common calcareous sandstones also occur. In thin section pressure welded contacts between the quartz grains are almost ubiquitous and the original character of the grains is therefore difficult to assess. Grains least affected by pressure solution appear angular.

Sandstones vary from fine to medium grained and are generally well sorted. In vertical profiles individual beds are of fairly constant grain size although rare graded units were observed. Bed thicknesses range up to 50 cms but most are less than 20 cms. Thicker units persist laterally for observed i.e. minimum distances of up to 60 m, thinner ones for much less. As they thin laterally beds often exhibit a reduction in grain size, tapering gradually into discontinuous bands and lenses of finer sandstone and may finally be reduced to silty laminae in the surrounding shales (Plate 46a).

The sharp lower surfaces of the sandstone beds often exhibit groove casts and other tool marks (Plate 46b) and a range of superbly preserved trace fossils (Plate 46c) including Kouphichnium, Rhizocorallium and (?)Palaeophycus. Other sandstones display convolute and loaded bases (Plate 47a). No fluting or larger furrows (e.g. Golding and Bridges, 1973) have been recorded.

Internally most sandstones exhibit planar lamination often associated with primary current lineation (Plate 47b). In several examples this passes laterally into ripple drift cross-lamination (Plate 47c) with both type "A" and "B" of Jopling and Walker (1968) being recognised. In many beds the planar lamination of lower parts gives way to an upper zone, up to 20 cms thick, of bidirectional cross lamination (Plate 48a), the knitted structures of de Raff et al, 1977. Upper surfaces of these sandstones commonly display symmetrical ripple marks. Cross bedding, often low angle, occurs sparingly in units up to 15 cms thick. These are often markedly lenticular and occasionally develop into rows of unconnected lenses. Planar laminated beds also pass laterally into flaser and linsen bedded areas (de Raff et al, 1977) which often show evidence of intense bioturbation with Planolites and Teichichnus (Plate 48b) abundant. Bioturbation of the only upper parts of sandstone beds in common and produces digitate tops. Bioturbation of thinner units leads to thorough mixing with the surrounding shales (Plate 46a).

Sandstone beds, largely free of the more obvious forms of bioturbation, often contain fine branching carbonaceous filaments extending down from their upper surfaces, almost certainly fossil rootlets. Shales interbedded with these units are often rich in plant debris and occasionally take on a clayey, seatearth-like texture and appearance.

In general, however, the shales are dark and silty and contain scattered ironstone nodules (Plate 46a). The sandstone plugged cylinders of Planolites are common and rare body fossils include bivalves, spinose Productid brachiopods, small gastropods and crinoid oosicles. Thin 2 to 3 cm thick sandstones packed with broken bivalve shells have been recorded and such coquinas often form the nucleus to lenticular calcareous nodules.

(ii) Interpretation

The sharp based sandstones with groove casts are the product of relatively sudden incursions of coarser sediment into a normally quiet, shale accumulating environment. Plane lamination and primary current lineation are commonly taken to indicate deposition under upper flow regime conditions (Allen, 1964 and Simons et al, 1965). The occurrence of ripple drift cross lamination indicates rapid fallout of sediment from suspension under tractional current conditions (Jopling and Walker, 1968). The development of convolute loaded bases to some sandstone beds is attributed to rapid sand deposition over thixotropic muds (Reineck and Singh, 1975).

The formation of flaser and linsen bedded zones is thought to indicate reworking of the sandstones by more protracted depositional processes generally referred to tidal currents (Reineck and Wunderlick, 1968) or wave action (de Raff et al, 1977). The abundance of symmetrical ripple marks on the tops of the sandstone beds and the occurrence of the complex knitted structures of de Raff et al point to the dominance of the latter process. The low angle cross bedded lenticular sandstones and rows of unconnected lenses are similarly interpreted as attenuated large scale wave ripples, though similar features elsewhere have been asigned a tidal origin (Johnson, 1975).

From a more general, environmental standpoint the abundance of trace fossils and the sparse shelly fauna indicate some marine influence.

(c) Lithofacies C

(i) Description

Quartz sandstones with variable carbonate content in the form of skeletal grains and cement constitute the principal lithologies of this lithofacies. Rock types range from pure quartz sandstones through to sandy, bioclastic limestones, with the bulk of the lithofacies falling into the category of calcareous sandstones.

Quartz sandstones with negligible carbonate content are fine to medium grained and well sorted. In thin section the grains display both pressure welding and syntaxial rim cements. Where the latter has preserved the original outlines of quartz grains they appear moderately well rounded. Pure quartz sandstones occur as irregular lenses and stringers within the more normal calcareous sandstones and occasionally form discrete beds up to 50 cms thick, but seldom without some calcareous patches. This intimate mixture of carbonate rich and carbonate free lithologies leads to honeycomb weathering, a characteristic feature of the lithofacies (Plate 48c).

In the more normal calcareous sandstones the quartz sand fraction is more poorly sorted ranging from fine to coarse grained and even small pebbles. Skeletal carbonate grains consist largely of crinoid ossicles although brachiopod, and coral fragments are common. In thin section, echinoderm plates exhibit syntaxial calcite overgrowths which poikilitically enclose adjacent quartz grains. Calcite/quartz contacts are typically highly corroded. Quartz grains are also observed penetrating skeletal grains, a product of pressure solution.

Skeletal grains may become so concentrated as to form lenses of coquina limestone (Plate 50c) containing only scattered quartz grains and pebbles.

The lithofacies displays a variety of sedimentary structures well seen in the Helaeth and Aber Sandstones. These include both trough and tabular cross bedding in sets up to 50 cms thick (Plate 49a) low angle and wedge shaped cross bedding (Plates 50b,65) planar lamination, cut and fill structures and wave ripple cross lamination. Current ripple cross lamination and ripple drift cross lamination (Plate 49b) also occur but are not common. Bedding planes often exhibit symmetrical wave ripples (Plate 50a) with wavelengths of up to 150 cms and amplitudes of up to 10 cms. Mound-like structures similar to those described by Goldring (1971) and Baldwin and Johnson (1977) also occur.

The lithofacies also contains a diagnostic suite of trace fossils characterised by vertical escape traces similar to Lockeia (formerly Pelecypodichnus) and protrusive Diplocraterion c.f. yoyo of Goldring (1971) (Plate 51). Other forms include Rhizocorallium, Phycoides and a variety of tracks and trails.

(ii) Interpretation

The range of sedimentary structures observed from the Lithofacies suggests deposition in an environment subject to a variable hydrodynamic regime in which both current and wave bedforms were generated. The lack of finer silt and mudgrade material indicates the predominance of higher energy tractional conditions and the efficient winnowing of such fines.

Trough and tabular cross-bedding, planar lamination, ripple and ripple drift cross lamination occur within the other lithofacies and their basic hydrodynamic interpretation has already been discussed. The large scale wave ripples recorded from the lithofacies are closely comparable with forms described by Campbell (1966). The origins of such structures are poorly understood, but perhaps reflect the larger orbital diameters of waves during storm conditions (Goldring and Bridges, 1973; Hames et al, 1975). The mound-like structures are compared with those described by Baldwin and Johnson (1977) are thought to indicate periodic emergence and gullying.

The often shelly nature of the Lithofacies and its contained trace fossil suiteindicate a depositional environment subject to strong marine influence.

.5 CHANNEL SANDSTONES

Throughout the succession channels incised through, and contemporary features of, palaeokarstic surfaces (Section 4.2) are filled by often complex siliciclastic sequences, here termed Channel Sandstones. In the Principal Area such units include the Benllech and Helaeth Sandstones, the coastal exposures of which were, in part, the inspiration for the present study. These and several of the minor channel sandstones exposed along this NE coast e.g. Aber and Dwlban Sandstones exhibit the necessary differentiation into distinct and recurrent lithofacies to allow detailed description and environmental interpretation. The Parc and Fedw Sandstones represent channel sandstones in the Penmon Area but are not described in detail. No such units have been recognised in the Straitside Area.

(a) Benllech Sandstone

(i) Section Details

Steep easterly facing cliffs between Traeth Bychan and Benllech provide superb sections in the Benllech Sandstone. In addition it is one of the few siliciclastic units exposed at the coast to outcrop significantly inland.

The coastal exposures show the Sandstone disposed in two discrete channels, a northern and southern one, incised into the limestones of the Dinas Beds (Fig.25). Channel form is particularly well demonstrated with contacts between the siliciclastic fills and flanking limestones well exposed at

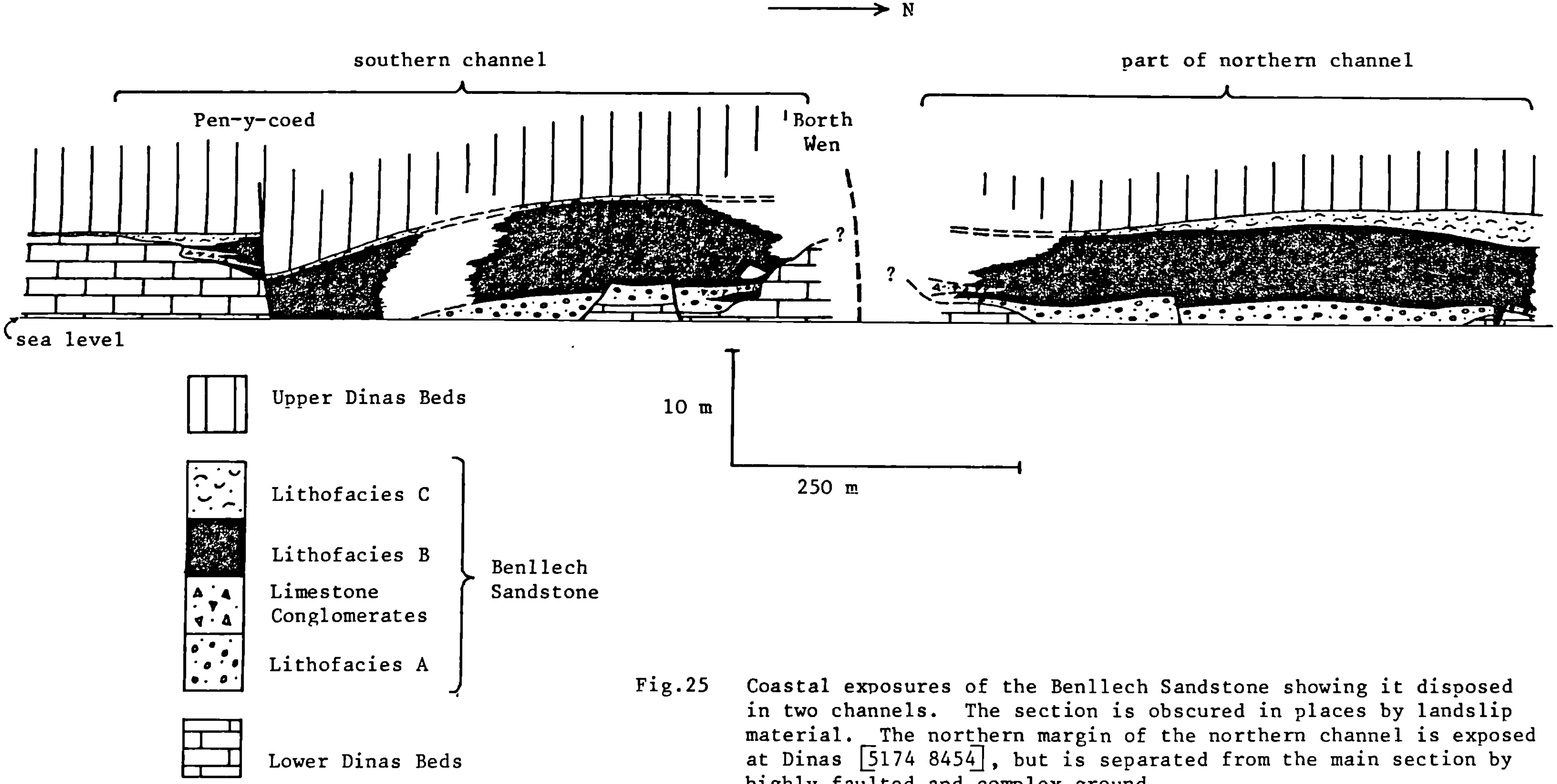


Fig.25 Coastal exposures of the Benllech Sandstone showing it disposed in two channels. The section is obscured in places by landslip material. The northern margin of the northern channel is exposed at Dinas $\boxed{5174\ 8454}$, but is separated from the main section by highly faulted and complex ground.

Pen-y-coed (Plate 5), Borth Wen (Plate 7) and Dinas. These exhibit complex relationships discussed and illustrated in detail in Chapter 3, Section 2c. The channel filling sequences and adjacent palaeokarstic surfaces are, for the entire length of the coastal outcrop, overlain by limestones of the Upper Dinas Beds.

The widespread outcrop of the Benllech Sandstone inland has, despite difficulties imposed by faulting and exposure, enabled channel distribution to be mapped. An anastomosing pattern is suggested with isolated limestone knolls of the Lower Dinas Beds, flanked on all sides by conglomerate plugged channels e.g. Vale of Cadarn (Fig.26).

Inland exposures exhibit little variation and are wholly developed within conglomerates and coarse pebbly sandstones, often cross bedded, of Lithofacies A. Lithofacies differentiation of the Benllech Sandstone and therefore detailed description and interpretation is restricted to the dual channel complex exposed at the coast.

Fig.25 illustrates the form and lithofacies distribution within these channels. The deepest parts of these channels are not observed, but inland exposures showing the Benllech Sandstone resting on the Upper Morcyn Beds demonstrate the complete removal of the Lower Dinas Beds in these areas and place a minimum depth to the channels of 19 m.

Both channels exhibit a similar sequence of lithofacies (Fig.25). They are floored by coarse conglomeratic units of Lithofacies A which pass upwards into sandstones and shales of Lithofacies B to produce a fining upwards sequence. This is

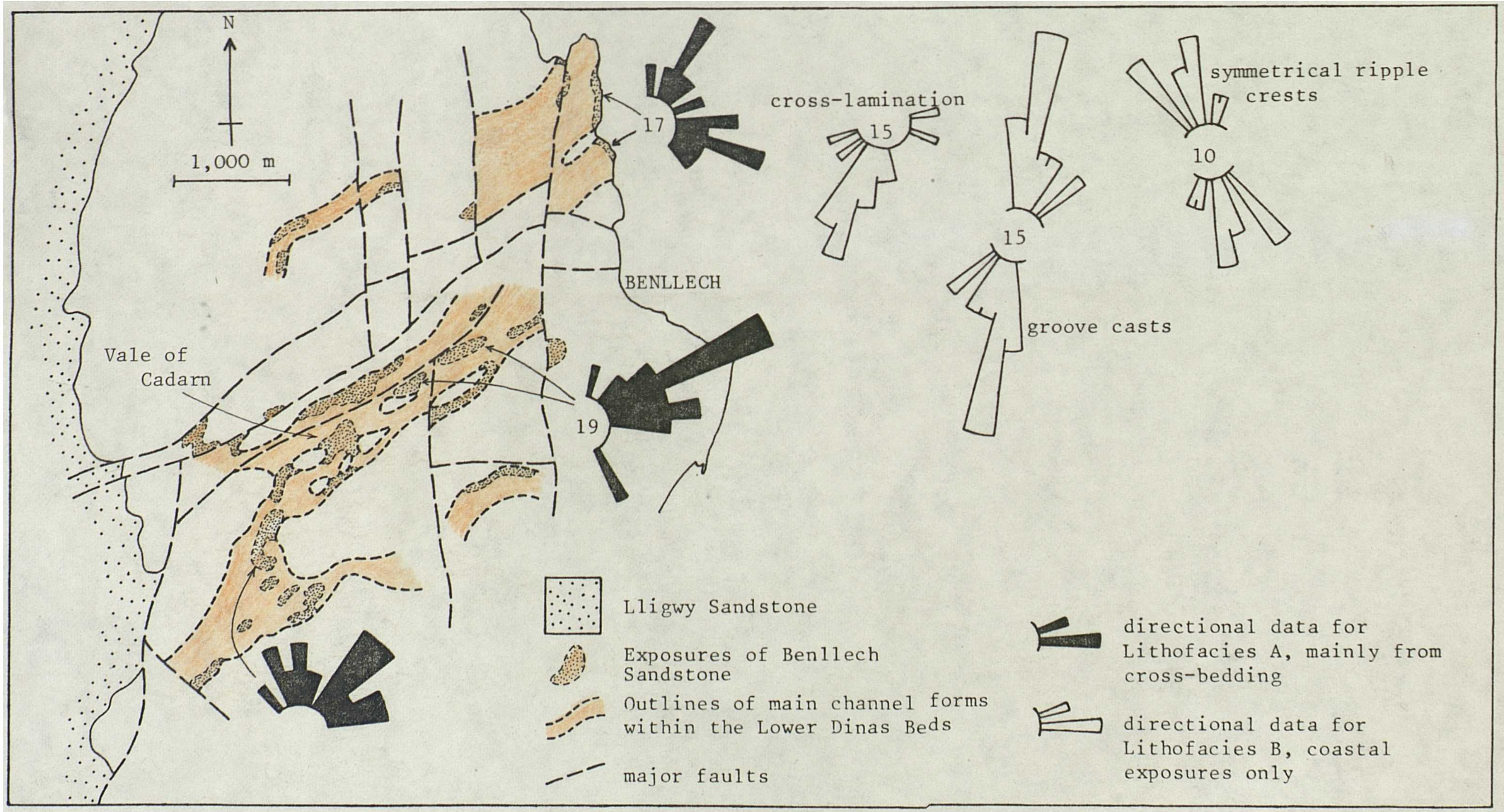


Fig.26 Mapped relationships of the Benllech Sandstone illustrating its distribution within channel complexes incised into the Lower Dinas Beds. Directional data for both the inland crops (Lithofacies A) and the coastal exposures (Lithofacies A and B) are also shown.

capped by calcareous sandstones of Lithofacies C, abruptly overlain by the fossiliferous shale which forms the base of the succeeding Upper Dinas Beds.

In the northern channel the conglomerates are for much of the length of the section only exposed between tides. They are largely massive in appearance although traces of cross bedding and low angle cross stratification are present. The unit is unusual here in exhibiting small scale horst and graben structures developed by a series of small faults (Plate 52a, b) (cf. Gill, 1979). These pass into the overlying sandstones and shales and die out as bedding plane slips. Most are probably the result of differential compaction of the channel fill sequence relative to the surrounding limestones after burial, but some appear to have affected the deposition of overlying units and a synsedimentary origin is indicated.

Conglomerates of the southern channel are exposed on its northern side near Borth Wen and are trough cross bedded with occasional reactivation surfaces (Plate 52c; cf. Johnson, 1975; McCabe and Jones, 1977). They occur in sets up to 2 m thick separated in places by irregular beds of dark grey argillaceous siltstone. These also pick out low angle bedding surfaces within the conglomerates (Plate 52c) and contain bands and thin beds of cross laminated sandstone. They also exhibit the trace fossils Teichichnus and Planolites and represent an intercalation of Lithofacies A and B.

At Borth Wen the conglomerates pass laterally and very rapidly across a narrow gully into medium grained, cross laminated sandstones. These apparently rest on shelves cut

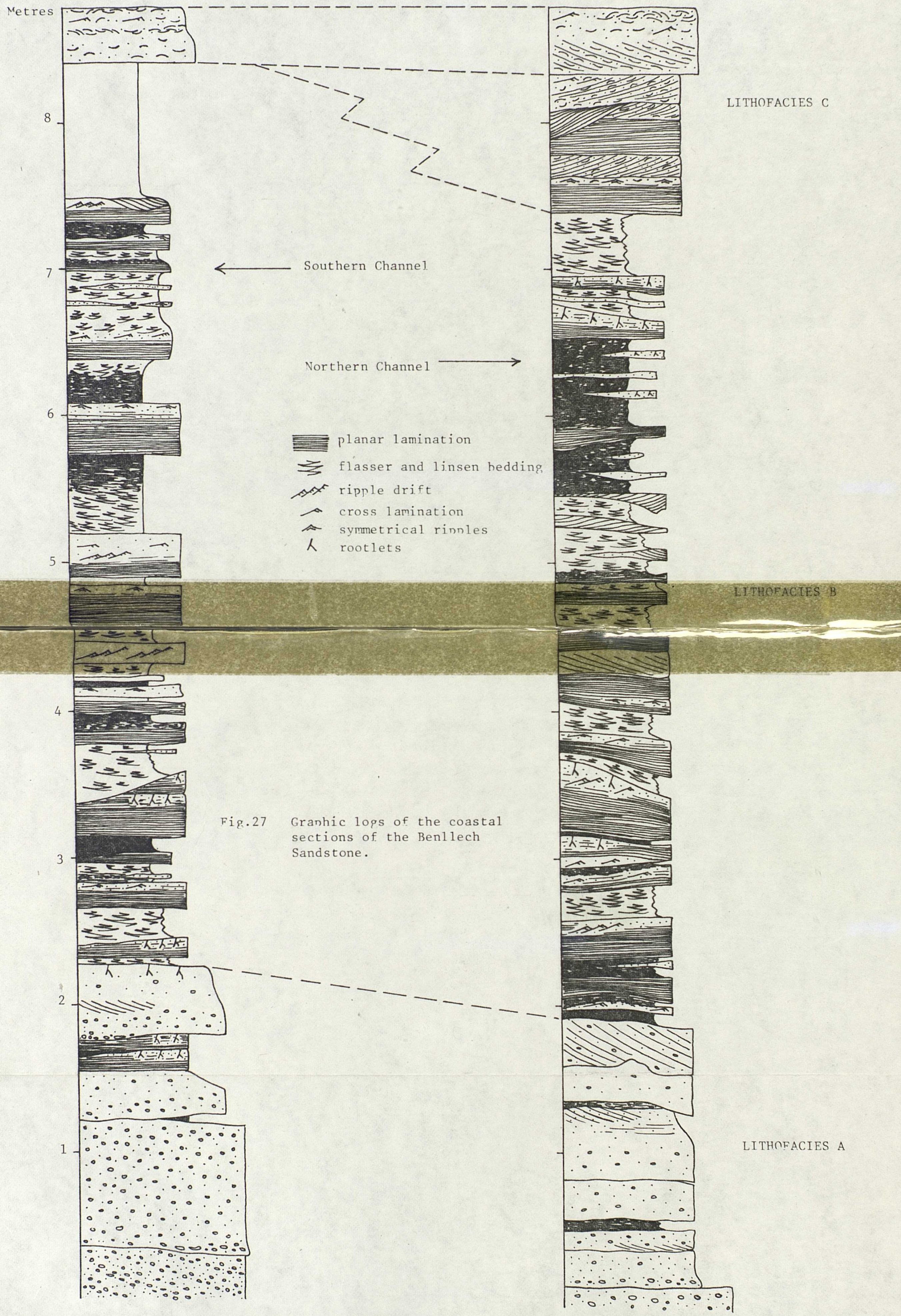


Fig.27 Graphic logs of the coastal sections of the Benllech Sandstone.

into the adjacent thin bedded limestones of the Dinas Beds which form the steeply rising northern bank of the channel. This is the "anomalous junction" described by Greenly (1919, pages 613 and 635) which he decided was unfaulted, a conclusion born out by careful examination of the exposure. It does however display further complexities with a mass of the Dinas limestones apparently having slumped down onto these marginal deposits (Plate 7), discussed in detail in Section 4.2c.

Palaeocurrent vectors derived from both the coastal and inland exposures of Lithofacies A display unimodal trends and on a larger scale indicate sediment transport paths east to north-east away from tracts of Ordovician and Precambrian rocks known to have formed an elevated hinterland during the deposition of the Anglesey Dinantian sequence (Fig.26) (Greenly, 1919 p.627, and George, 1974 p.112).

Towards the northern end of the northern channel the conglomerates of Lithofacies A are abruptly overlain by sandstones and shales of Lithofacies B. Towards the southern side of the channel a more gradational passage occurs with beds of coarse pebbly sandstone, and then medium grained sandstones still with scattered quartz pebbles forming a transitional unit. These beds, which also contain plant fragments and rootlets, exhibit tabular cross bedding indicating transport directions the same as the underlying conglomerates i.e. easterly.

In the southern channel, the transition to the sandstones and shales of Lithofacies B is less simple (an intercalation of the two facies at Borth Wen has already been noted). Conglomerates pass upwards into coarse pebbly sandstones and medium grained sandstones as above, but the upper units of this transition, well sorted medium grained sandstones, exhibit rapid lateral changes in thickness, splitting, and a set of distinctive erosion surfaces. These features are illustrated in Plates 53 and 54. As can be seen there are two groups of these truncated and wedge shaped beds. Individual sandstone wedges exhibit planar to low angle cross stratification, and some display groove casted bases. Tops are planar or rarely display "rib and furrow" structure, others have simple pit like traces. The curved erosion surfaces which truncate the lamination of these beds (Plate 54b) have sinuous tracks, probably Gyrichnites preserved on them (Plate 54c) and occasionally have adhering shale clasts. Both "sets" of these units rest on bioturbated sandstones with abundant Rhizocorallium, and similarly pass laterally into highly bioturbated zones in which Teichichnus forms the dominant and ubiquitous form. This zone is also of limited lateral extent and in turn gives way to more readily recognisable units of Lithofacies B (Plate 53c).

The coastal exposures of the Benllech Sanstones provide the "type locality" for the sandstones and shales of Lithofacies B and exhibit in detail the features and characteristics described in Section 4.4b. Towards the northern end of the northern channel the lower units of the lithofacies exhibit convolute and loaded bases (Plate 47a) and it is in this area where deposition of the sandstone beds appears to have been partly controlled by the synsedimentary faults described above. Towards the southern limits of exposure in the northern channel the sandstones of the lithofacies collectively thin or die out to produce a shale dominated unit with abundant ironstone nodules. A similar style is observed in the southern channel where sandstone beds as they are traced southwards over the wedge shaped transitional units give way to a shale rich area. This appears to extend from the large landslip (Plate 53a) for the remainder of the channel to the southern margin at Pen-y-coed. Within it ironstone nodules are again conspicuous and spinose productid brachiopods and the byssate bivalve Sanguiolites are more common than in the normal sandstone-rich parts of the lithofacies. Other features of the lithofacies in this southern channel include injectional structures (Plate 45b) and shallow scours incised through sandstone beds (Plate 47b) which have not been observed in the northern channel sequence. Micritic limestone nodules which weather to a distinctive cream colour amongst the browns and greys of the sandstones and shales also occur in the southern channel. They occasionally coalesce to form more continuous beds and may contain bivalve fragments.

A characteristic feature of the lithofacies, well seen in both channels, is the rapid lateral and vertical passage of sharp based sandstone beds often with traces of rootlets into flaser and linsen bedded zones often intensely bioturbated with abundant Teichichnus.

Towards the top of the sandstone/shale sequence in the northern channel, where it is accessible to the south of the section, shales beneath the succeeding units of Lithofacies C take on a seartearth like appearance with abundant plant debris. In the southern channel, however, shales at the same level contain the marine bivalve Dunbarella and small (possibly stunted) rhynchonellid brachiopods. The higher sandstone beds of Lithofacies B, in both channels, are often slightly calcareous.

Current vectors for the lithofacies derived from groove casts, ripple drift cross lamination etc. are shown in Fig.26 and indicate a northerly derivation.

Calcareous sandstones of Lithofacies C cap both channel sequences and extend beyond onto the adjacent palaeokarstic surfaces against which they gradually pinch out. The lithofacies attains its thickest ~~development of~~ development of 6 m at the northern side of the northern channel (Plate 55). It is largely inaccessible here, but can be examined in fallen blocks and appears to include beds of sandy coquinoid limestone amongst the more normal trough and low angle cross bedded, calcareous sandstones.

The unit gradually thins southwards so at the southern end of the northern section it is less than 2 m thick (Fig.27). Here trough and low angle cross bedded units are overlain by a distinctive 50 cms bed of shelly sandstone and it is this same bed, reduced to 35 cms (Fig.27) which forms the only representative of the lithofacies in the southern channel sequence.

It is convenient to discuss the depositional models for the Benllech Sandstone in terms of environmental interpretations for individual lithofacies. These should be studied in conjunction with the lithofacies descriptions and basic hydrodynamic interpretations presented in Section 4.4.

(ii) Environmental Interpretation : Lithofacies A (see also Section 4.4a)

It is clear from stratigraphic relationships that the channels within which the Benllech Sandstone is confined were contemporary features of the palaeokarstic surfaces through which they are incised. They were formed, therefore, during emergent episodes and demonstrate the dissection of the limestone platform in response to a lowered base level (Section 3.2c).

Within this stratigraphic context, and given the terrestrial setting suggested above (Section 4.4a), several factors combine to make a fluvial origin for Lithofacies A unequivocal: the basal position within channel filling sequences; the presence of tractional sedimentary structures and unimodal palaeocurrent trends; the provenance of the pebbles and transport paths away from contemporary upland areas; and the coarse poorly sorted nature of the deposits. Poor pebble segregation within pebbly sandstone is also thought to be indicative of fluvially deposited as opposed to wave reworked gravels (Clifton, 1973).

Fluvial systems which transport mainly gravel are generally braided in character (Harms et al, 1975) although some low sinuosity meandering streams carrying sand and gravel (McGowen and Garner, 1970) or exclusively gravel (Gustavson, 1978) also exist. The often rapid lateral variations in grain size the lithofacies exhibits and the occurrence of siltstone laminae are indicative of a variable flow regime also suggestive of braided streams in which both fluctuations in discharge and rapid shifting of bars and channels leads to such conditions (Smith, 1970). This is further supported by the frequent erosion of the siltstones and their incorporation as clasts within the conglomerates (Allen, 1970). The general lack of thicker accumulations of fine sediment attributable to an overbank or abandoned channel fill origins (but see the Helaeth Sandstone, Section 4.7a) is in accord with a braided model (Allen, 1965; Smith, 1970 and Harms et al, 1975), although coarse grained low sinuousity meandering systems may be similarly impoverished in fines (Moody-Stuart, 1966 and McGowen and Garner, 1970).

The often massive appearance and occasional horizontal stratification of the lithofacies is consistent with deposition by pebbly braided streams where longitudinal bars, the most common bar form (Williams and Rust, 1969 and Smith, 1970) are known to develop similar structureless or crudely stratified deposits

(Rust, 1972a and Smith, 1974). These also exhibit the same a (t) b (i) pebble imbrication fabrics as recorded from the massive parts of Lithofacies A (Rust, 1972b and Harms et al, 1975).

Tabular cross bedding in pebbly braided stream deposits is generally attributed to transverse bars (Smith, 1970 and 1974) although longitudinal bars may also develop downstream slip facies and generate identical structures (Costello and Walker, 1972; Smith, 1974 and Harms et al, 1975). Where tabular cross bedding occurs within more massive units the latter origin seems more likely. Trough cross bedding has a more diverse origin. It may be developed by the migration of either dunes within the deeper, more active parts of channels or sinuous crested transverse bars (Smith, 1974). Successive fills of scour troughs described from low sinuosity streams produce trough cross bedded deposits (McGowen and Garner, 1972) as do a host of more ephemeral situations related to falling stage modification of braid bars (Williams and Rust, 1969 and Rust, 1972a).

It is important to appreciate the scale on which the above braided or low sinuousity fluvial systems were operating. The major channel features, incised through any one palaeokarstic surface , were but single anabranches in a larger anastomosing pattern (Fig.18) (The term anastomosing is adapted from the usage of Schumm, 1971). Units of Lithofacies A were deposited in the <u>lower</u> portions of <u>each</u> anabranche. They represent the products of fluvial braid plains established in each of these individual channels.

(iii) <u>Environmental Interpretation : Lithofacies B</u> (see also Section 4.4b)

A marine influence on the depositional environment for

Lithofacies B was indicated in Section 5.4b. The occurrence of the Lithofacies as part of channel-fill sequences, overlying fluvial deposits, leads to an interpretation for units of Lithofacies B as representing estuarine deposits. The Benllech Sandstone channels were developed during a regressive episode when the Dinantian shelf was emergent and fluvial incision to a lowered base level could take place. The establishment of marine conditions within these channels, as indicated by units of Lithofacies B, must, therefore, result from the rising sea level of the ensuing marine transgression which culminated in a return to carbonate deposition on the shelf. The term estuarine as it is applied to the channel sequences is, therefore, used in the sense of a drowned river valley rather than its more common application to the tidally influenced mouths of rivers.

The rarity of shelly fossils, particularly brachiopods, may be a function of reduced salinity or high turbidity of the water (Raup and Stanley, 1971) both equally consistent with an estuarine setting. So too is the occurrence of the trace fossil *Kouphichnium*, attributable to the walking tracks of Limulids (Hantzschel,1975) which abound in the present day estuaries of the eastern seaboard of America. A comparable interdistributary bay environment has been given to a similar association of sharp based sandstones and *Kouphichnium* from the Namurian (Collinson and Banks, 1975).

The rapid lateral passage of bioturbated flaser and linsen bedded areas into root infested sandstone beds and seatearth like shales indicates an environment in which low emergent banks, colonised by plants, were surrounded by areas of shallow water. Again this would seem in keeping with an estuarine flats situation.

Whilst the estuarine conditions envisaged for units of Lithofacies B resulted from drowning of river valleys once they have formed, such estuaries are subject to the same forces and processes which operate within their tidally classified relatives. These are subdivided on the basis of tidal range and influence into micro-, meso- and macro-tidal (Hayes, 1976).

The higher tidal range in meso- and macro-tidal estuaries i.e. over 2 m, leads to a dominance of current produced bed forms; ripples, dunes and tidal deltas abound (Hayes, 1976 and Boothroyd and Hubbard, 1976). Marginal mud flats develop meandering tidal channels (Hayes, 1976). In Lithofacies B the paucity of cross bedding and the total lack of any of "herringbone type", combined with an absence of fining upwards sequences attributable to meandering tidal channels (Chisholm, 1970) favour a micro-tidal interpretation. Modern micro-tidal estuaries of the North West Gulf of Mexico described by McGowan and Scott (1976), formed during the Flandrian transgression, appear a good analogue for those developed in the Anglesey Lower Carboniferous.

The main agents which affect deposition in such micro-tidal estuaries are waves, rivers and storms with only minor tidal influence (Hayes, 1976).

In units of Lithofacies B the abundance of symmetrical ripple marks, of flaser and linsen bedding and of bi-directional cross lamination (de Raff et al, 1977) demonstrate the more or less constant effects of wave activity.

Intercalation within channel-fill sequences of Lithofacies B with Lithofacies A reflects, not unexpectedly, strong fluvial influences on estuarine environments as they first become established, i.e. as the channels are first drowned. Individual units of trough cross bedded conglomerate, separated by bioturbated silty shales, (Plate 52c) are likely to represent the deposits of single powerful flood events (cf. Bluck, 1967). These swept sediment down the major channels and deposited sheet-like beds of gravel as they debouched onto newly formed estuarine flats (cf. McGowen and Scott, 1976). As transgressions advanced and base levels rose, the potency of river floods diminished and deposition, certainly of the coarser fluvial material, was excluded from the more distal estuarine environments.

In channel sandstones where such intercalation of the two lithofacies is not apparent more rapid transgression, or deposition in less fluviall active anabranches of channel systems may be indicated.

The sharp based sheet sandstone of Lithofacies B have been interpreted as marking short lived, rapidly deposited, high energy events (Section 4.4b). Where these sandstones immediately overlie units of Lithofacies A it is again logical to evoke an origin related to river flood events. In fluvial sequences such sheet sandstones attributed to such events and commonly explained as crevasse splay deposits.

Generally a feature of meandering rivers or delta distributaries, crevasse splays are produced when rivers burst through their confining levees during periods of flood. A lobate, but often extensive sheet of sand is deposited. Ancient sheet sandstone assigned to this mechanism exhibit many of features of the Lithofacies B sandstones: sharp erosive bases, sole marks, planar and ripple drift cross-lamination, occasional grading and rootlets (Collinson, 1978 and Elliott, 1978). Crevasse splays into inter-distributary bays on deltas may also display wave rippled tops and bioturbation effects (Elliott, 1974) and the similarities of these to the sandstones of Lithofacies B is striking. Ancient examples of this latter variety have been described by Collinson and Banks (1975) from Namurian deltaic sequences.

In the Benllech Sandstone however there is no evidence of contemporary fluvial channels adjacent to the estuarine sequences and a true crevasse splay origin is therefore unlikely. If the sheet sandstones of Lithofacies B are indeed comparable and result from river flood events they are derived from up-channel fluvial sources and may represent the distal equivalents of the gravel sheets described above.

Anomalous current vectors, at high angle to those recorded from underlying fluvial lithofacies, and the occasional calcareous nature of the sheet sandstones relating them more closely to Lithofacies C, prompted an investigation of other processes likely to emplace sheet sands into estuarine environments. Storms were the other obvious candidate.

Ancient sandstone beds attributed to storm activity are generally laterally persistent sheet sandstones (Goldring and Bridges, 1973; Johnson, 1978 and Vos, 1977) in accord with the geometry of present day storm generated deposits both of washover type (McGowen and Scott, 1976) and from the sublittoral shelf environment (Reineck and Singh, 1972).

The internal structures of storm generated deposits are produced by a combination of tractional currents and storm wave activity. Tractional currents develop in response to storm surge or storm surge ebb and may be superimposed upon day to day tidal and rip current systems (Johnson, 1978). Deposition from short lived storm induced tractional currents can give rise to graded and waning flow "Bouma" type sequences (Banks, 1973), but more often tends to produce units which are largely planar or ripple drift cross laminated (Goldring and Bridges, 1973 and Vos, 1979). Low angle cross bedding has been interpreted in terms of Simons et al's (1965) washed out dune phase (Johnson, 1978).

Deposition of storm suspended sediments under diminishing storm wave conditions may also produce planar laminated units (Reineck and Singh, 1972) or may give rise to undulatory cross stratification thought to represent large scale wave cross lamination (Campbell, 1966 and Goldring and Bridges, 1973). This appears identical to the humocky cross stratification of Harms et al (1975), Vos (1979) and Cant (1980). More potent or prolonged reworking by wave activity may produce the more complex knitted structures of de Raff et al (1977). Wave rippled tops are ubiquitous and indicate post storm reworking.

Trace fossils restricted to bounding surfaces or bioturbation of only the upper portions of sheet sands are strongly indicative of rapid deposition under storm conditions (Goldring and Bridges, 1973).

Clearly there are many similarities with the Lithofacies B sandstones. Of the two types of storm sheet sands, sublittoral shelf and washover, the latter is favoured. The low diversity bivalve dominated shelly fauna recorded from the estuarine shales of the Benllech Sandstone indicates a restricted environment and suggests the presence of off-shore barrier complexes. Under transgressive regimes such barriers migrate landwards. Sediment is transferred from the seaward to the landward side of the barriers and this is achieved principally by a process of storm washover (Swift et al, 1971; Kraft et al, 1973). In the micro-tidal, back barrier estuaries of the North West Gulf of Mexico washover sheet sands constitute a major feature (McGowen and Scott, 1976). Other ancient sheet sandstones assigned to a washover origin have been described by Johnson (1975), Bridges (1976) and Vos (1977).

Both mechanisms, which have been discussed, river flood and storm washover, may have operated within the Dinantian estuaries of the Benllech Sandstone (Fig.28). Indeed the two processes are in a sense related since storm events may generate barrier washovers whilst associated heavy rainfall causes river flooding (McGowen and Scott, 1976).

The sets of wedge shaped beds observed in the southern channel at the junction between Lithofacies A and B are not well understood (Plates 53,54). The planar laminated and groove casted nature of the beds firmly links them with the sharp based sandstones of Lithofacies B. Prior to truncation by their distinctive erosion surfaces therefore it seems likely that these beds were also laterally extensive and emplaced by short lived, high energy events related to storms or river flood. Clearly, however, these particular units were deposited in a setting where they were prone to subsequent erosion. Similar structures have been described by Goldring (1971) from Devonian shallow marine deposits and termed 'erosion ripples'. These were thought to result principally from the effects of wave action in shallow water. Comparable Recent structures have been described from lacustrine beach deposits (Davis et al, 1972). Here curved erosion surfaces, formed during storm reworking, truncate earlier planar and low angle swash lamination within beach ridges. The Benllech Sandstone units may therefore record similar reworking. It is perhaps possible that these beds were deposited over an earlier abandoned fluvial bar or a prominance in the channel's limestone floor so that with storm surge ebb or receding flood waters they formed possibly emergent sandy shoals within the then only recently formed Benllech Sandstone estuary. The flanks of such sandbodies would have been attacked and slowly consumed by daily wave action and/or estuarine currents leaving the remnants now observed. With rising sea level the inherited effects of such underlying prominances would be lost and the overlying sheet sandstones of Lithofacies B were deposited without subsequent modification.

(iv) Environmental Interpretation : Lithofacies C (see also Section 4.4c)

Lithofacies C displays in detail many of the features observed from the deposits of modern beach and upper shoreface environments. These are typified by moderately well sorted quartz sands with a variable content of skeletal grains (Thompson, 1937). Lenses and stringers of coarser shell debris and larger pebbles are also common in such units (Clifton et al, 1971; Clifton, 1973 and Reineck and Singh, 1975). The sedimentary structures developed by these environments and the processes which form them have been described by Clifton et al (1971), Howard and Reineck (1972) and Davidson-Arnott and Greenwood (1976) and bear a remarkable resemblance to thos observed in Lithofacies C. Ancient sediments which appear very similar to Lithofacies C have also been assigned to these environments (Campbell, 1971; Johnson, 1975 and Harms et al, 1975).

Low angle wedge shaped cross stratification and planar lamination is attributable to the beach swash zone. Trough cross bedded units may be developed by migrating megaripples in beach ridge and runnel systems (Clifton et al, 1971) or in the crest facies of near shore bars (Davidson-Arnott and Greenwood, 1976).

The characteristic vertical escape traces and protrusive Diplocraterion are members of Seilacher's (1967) Skolithos Facies indicative of an environment where the substrate is unstable and prone to both rapid erosion and deposition. This is compatible with the beach and nearshore setting envisaged for Lithofacies C and indeed very similar traces have been observed from modern equivalents (Howard, 1972).

In view of the discussion for Lithofacies B it appears likely that the beach deposits represented by Lithofacies C accumulated in a micro-tidal setting. Beaches are in fact common features of micro-tidal coasts where they develop in response to wave and/or storm activity (Davidson-Arnott and Greenwood, 1976; Fraser and Hester, 1977).

Within the context of the channel fill sequences, the capping nature of the lithofacies and its often sharp erosive contact with underlying units of Lithofacies B indicates a transgressive origin. They are interpreted as transgressive barrier spit deposits (cf. Bridges, 1975). Storm washover is an integral feature of transgressive barrier systems and the above discussion for Lithofacies B is therefore in accord with such a model.

The abrupt contact of units of Lithofacies C with the overlying Upper Dinas Beds records the final drowning of the Benllech Sandstone channel complex, a conclusion bourne out by the coincidence of this level with that of adjacent palaeokarstic surfaces. Thus as rising sea level reached the level of the palaeokarstic highs between channels rapid inundation of the shelf took place and active carbonate deposition was resumed.

Such contacts represent the 'Kick-back' transgressions of Irwin (1965) and are further discussed in Chapter 5, Section 6b.

(b) Helaeth Sandstone

The Helaeth Sandstone, exposed in the cliffs of Porth Helaeth, occurs at the base of the Moelfre Limestone Formation. Fig.12 shows the extent of outcrop and the main cliff sections. There are no inland exposures.

The channel form of the Helaeth Sandstone is not well displayed. Only its western margin is exposed. The gradual bevelling of the underlying Royal Charter Beds points to the easterly deepening hollow which the Sandstone is thought to occupy.

The Helaeth Sandstone is readily divisible both geographically and in terms of lithofacies distribution into two sections (Plate 56). The low cliffs to the west, beyond the creek which marks a small fault, are developed in conglomeratic lithofacies. To the east, the remaining lithofacies of the Sandstone are exposed for 300 m at the base of northerly facing cliffs before low south easterly dips bring the limestones of the overlying Helaeth Beds down to low water mark.

(i) Description of Western Exposures : Conglomerates and related lithofacies

Plate 56 illustrates a detailed section through the conglomerates and related lithofacies which form the western cliffs. These are divisible into a lower and upper unit.

The lower unit to the west as it emerges from boulder clay consists of a medium grained quartz sandstone bed overlain by variegated red and green silty shales (Plate 57). The sandstone forms a basal bed resting on the upper surface of the Royal Charter Beds which here appears quite smooth. The shales contain simple cylindrical burrows and possible desiccation cracks. Thin sandstones thicken as the shales quickly give way to dipping, crudely sigmoidal, lenticular sandstone beds (Plate 57). These appear massive and have sharp, but not strongly erosive bases. Both tops and bases exhibit simple tracks and trails. The lenses dip by approximately 15° to the east-south-east (110° to 120° to grid north) and with regional dips at 2° to 4° to the south-east it is clear that the dips in the sandstones are largely depositional. The upper surface of the basal sandstone at this point acquires a similar dip, but also displays a series broad trough features aligned perpendicular to the dip, and internally faint traces of cross bedding directed up the dip are visible (Plate 58a, Fig.29). Poorly developed linguoid ripples occur on the upper surface and indicate currents flowing obliquely or perpendicular to the direction of dip.

To the east the sandstones become more pebbly, indeed individual sandstone beds become more pebbly down dip. Silty shale intercalations are also reduced. These pebbly sandstones probably pass laterally into the adjacent dipping beds of the conglomeratic lithofacies although the actual transition is obscured by boulders.

The conglomerates are of the normal type, Lithofacies A, and are characteristically gradational with coarse pebbly sandstones. The dipping units developed in this lithofacies are more regular than the preceding lenticular sandstones and occur as parallel sided beds up to 1.80 m thick. They are separated by thin, up to

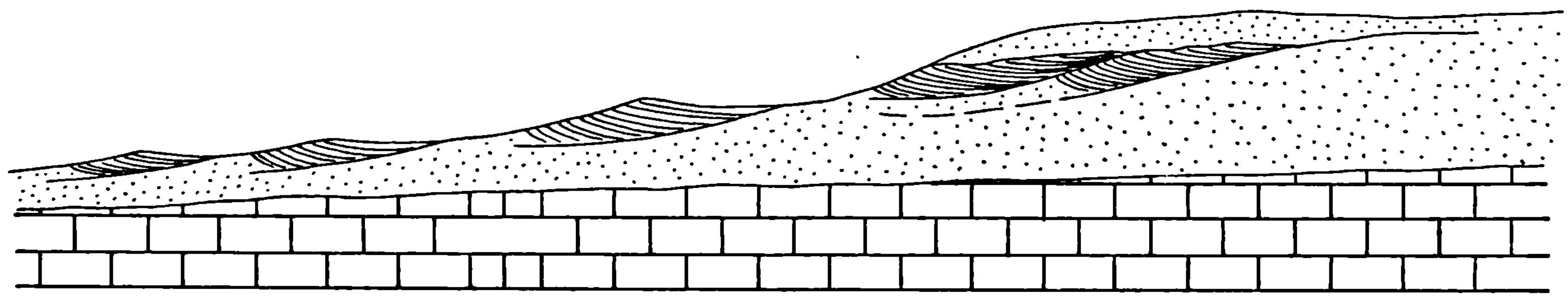


Fig.29 Extreme western end of the western exposures of the Helaeth Sandstone; basal pebbly sandstone with scalloped top and updip cross bedding resting on bevelled top of Royal Charter Beds.

5 cms thick, units of silty shale. Individual beds may be wholly conglomeratic, but many consist of a coarse pebbly sandstone lower portion and a conglomeratic upper part. The contact between these is generally fairly sharp (Plate 58b) but in some cases a more gradational coarsening upwards is indicated. Internally the beds appear mainly massive although traces of cross bedding and cut and fill structures are observed. These show a variable orientation, generally oblique or perpendicular to the depositional dip, but occasionally directed down dip. The upper truncation surface to the lower unit also has discrete conglomerate filled scours associated with it (Plate 58c).

The dipping units pinch out against the underlying limestone surface which in this easterly part of the section displays broad pot-holes filled by congomerate (Plate 9a,b). Depressions within the overlying conglomerates (Plate 9c) indicate the continued deepening of these features after burial (Section 3.2c).

These shallowly dipping conglomerates and coarse pebbly sandstone beds continue for the rest of the section before they disappear eastwards beneath low water mark and southwards below the single beach of Porth Helaeth.

This lower unit is overlain by an upper conglomeratic unit also attributable to Lithofacies A (Plate 56). This has a sharp erosive base and is trough cross bedded throughout (Plate 45a). Current directions to the north-east (Fig.30) are indicated. 4 to 5 m of this unit are exposed before the outcrop becomes obscured by boulder clay. The basal limestone of the Helaeth Beds, a conspicuous bed with dolomitised burrows, is observed after a break of 2 m in the section.

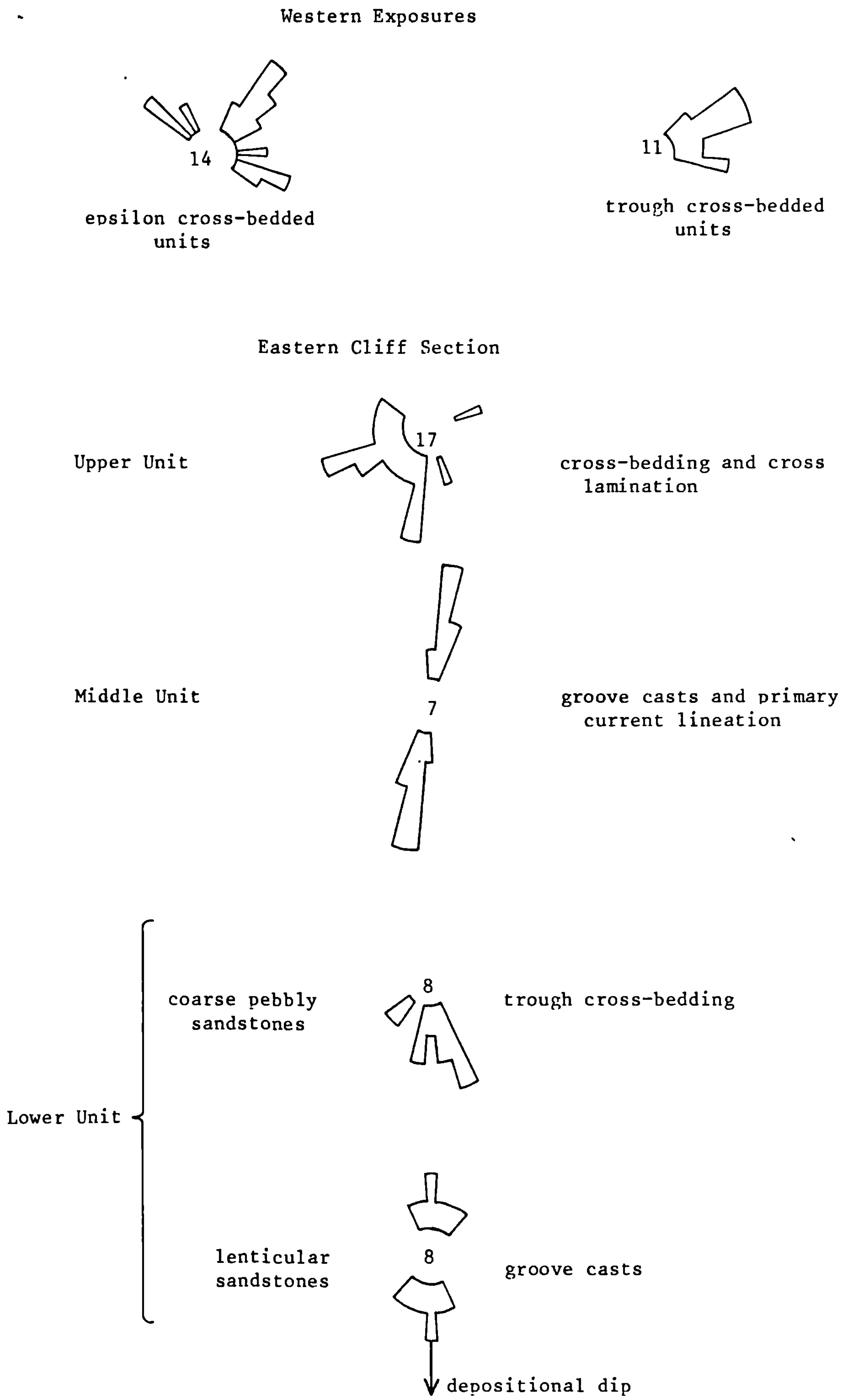


Fig.30 Directional data from the Helaeth Sandstones

(ii) Environmental Interpretation for the Western Exposures of the Helaeth Sandstone

The fluvial affinities of the conglomeratic lithofacies A has already been outlined (Section 4.5) and it is clear from their close association that the shales and lenticular sandstones of the lower unit belong to the same setting.

These lenticular sandstones exhibit the following features, shallow depositional dip, shaly intercalations, up dip passage into shales, increasing pebble content down dip and current directions sub-parallel to their strike, all indicating that these dipping beds are a form of epsilon cross bedding (Allen, 1963 and 1965a) formed by a process of lateral accretion. So too, presumably, must the contiguous conglomeratic units to the west of the section which display the same shallow dip and shaly intercalations, yet these also exhibit many features incompatible with the classical idea of epsilon cross bedding (Allen, 1965). They rarely fine up dip indeed they sometimes coarsen, and whilst scours are often orientated perpendicular to dip, cross bedding is occasionally directed down it! These differences between the lenticular sandstones and the conglomeratic units and the latter's variation from the norm appear related to the coarsening grain size and are thought to be indicative of a changing fluvial regime.

Epsilon cross bedded units within fluvial sediments are generally interpreted as point bar deposits of meandering streams (Allen, 1963 and 1965a). Their formation and preservation are thought to indicate a variable discharge (Collinson, 1978), possibly seasonal. Descriptions of modern point bars and meandering streams are largely confined to sandy fluvial systems (Sundborg, 1956; Jackson, 1976). Similarly, ancient fluvial sequences in which epsilon cross bedding has been observed are generally sandy deposits (Allen, 1965a; Moody-Stuart, 1966; Puigdefabregas, 1973; Leeder, 1975 and Nami, 1976).

The shales and lenticular sandstones to the east of the lower unit display features readily comparable with modern sandy point bars. The cross bedding directed updip and trough features of the basal sandstone bed (Plate 58; Fig.29) are reminiscent of scroll bar development and the formation, on many point bars, of a ridge and swale topography (Sundborg, 1956). In a meandering stream, scroll bars originate in the deeper parts of the channel and them migrate towards and onto the point bar surface. With slip faces often directed up the depositional ramp of the point bar it is clear how the up dip cross bedding of the Helaeth example could be produced. The associated trough features correspond to the swales developed between successive scroll bar ridges in which fine sediment may accumulate. Similar features have been described from Miocene point bar deposits (c.f. Puigdefabregas, 1973).

Where scroll bars attain a position high up on the point bar surface and deposition of finer sediment from suspension becomes important they form a series of elongate, dipping, lenticular sand bodies with finer grained intercalations (Sundborg, 1956). This description bears a remarkable resemblance to that for the dipping crudely sigmoidal sandstone lenses from the Helaeth Sandstone. The thicker shales associated with these units to the east of the section are interpreted as overbank fines.

The shallowly dipping conglomeratic units to the east of the section are thought to represent a gravelly extension of these point bar units. In Section 4.5a it was concluded that the characteristics of the conglomerates and pebbly sandstones of Lithofacies A were compatible with their deposition by pebbly braided streams. These, however, often develop meandering reaches and point bars (Williams and Rust, 1969 and Rust, 1972a). Smith (1974) briefly aludes to deposition by lateral accretion on such point bars. Analogous features, laterally accreting gravel meander lobes, have also been described from low sinuosity rivers (Gustavson, 1978). Gravelly point bars developed within a dominantly braided stream system exhibit many of the features of their associated braid bars and this may explain the divergence of the conglomeratic units from the normal style of epsilon cross bedding. This is a function of the more powerful currents, particularly during periods of flood, which operate within a gravel transporting system. Thus, coarser grain sizes may be carried onto the upper parts of the point bar surface and prevent the more normal fining upwards trend from developing. Indeed with many gravelly bars, including point bars, exhibiting a downstream fining (Smith, 1974), downstream migration may give rise to the coarsening upwards styles noted from the Helaeth units. The same phenomena may also result from rising discharge and increasing competency of river currents during flood (Eynon and Walker, 1974). The upper surfaces of gravel point bars are often dissected by a series of shallow chute channels terminating in chute bars (McGowen and Garner, 1970 and Gustavson, 1978). Superimposed sinuous crested transverse bars may also extend across the point bar surface (Smith, 1974 and Gustavson, 1978). These features could, especially with falling stage modification, give rise to the anomalous cross bedding directions noted from the conglomeratic beds. The conglomerate filled scours observed from

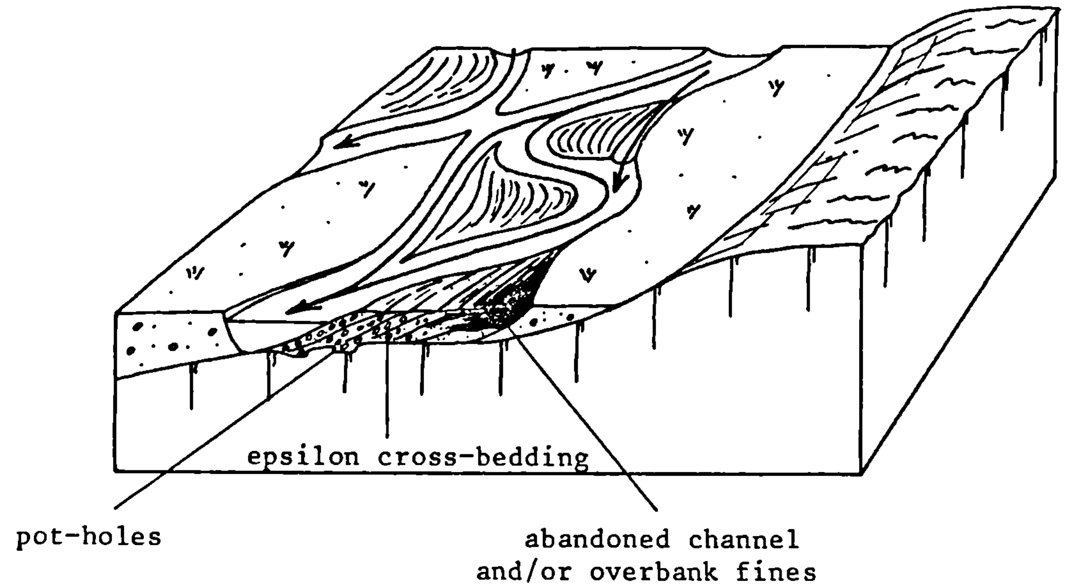


Fig.31 Environmental reconstruction for the western exposures of the Helaeth Sandstone : a possible model for the formation of the observed 'coarsening-sideways' sequence.

these units may represent small chute channels or may be developed by dissection of the bar surface during falling stage (Williams and Rust, 1969). During such periods the finer silts which separate many of the dipping beds were deposited from suspension.

The lateral passage from shales and sandstones into conglomeratic units represents a "coarsening sideways" sequence. Coarsening upwards sequences from Pleistocene outwash gravels have been interpreted as resulting from the reactivation of temporarily abandoned braided river channels (Costello and Walker, 1972). These units of the Helaeth Sandstone are thought to have been formed by a similar mechanism, but where the dominant process was one of lateral rather than vertical accretion. Thus reduced flow through a less active side channel allowed an embrynonic sandy point bar to develop with a high proportion of fine sediment deposited from suspension. The conglomerates mark the influx of gravels under more active channel conditions, but indicate the retention of a point bar mode of deposition (Fig.31).

The overlying erosively based, trough cross bedded unit represents a subsequent braided channel deposit with sinuous crested transverse bars (Smith, 1970 and 1974).

(iii) Description of Eastern Cliffs : Calcareous Sandstones and related lithofacies

Figs.32,33 illustrate the ready division of the eastern cliff section into three units, a lower unit characterised by dipping lenses of sandstone, a thin middle unit in the shales and sandstones of Lithofacies B and an upper unit developed in the calcareous sandstones of Lithofacies C.

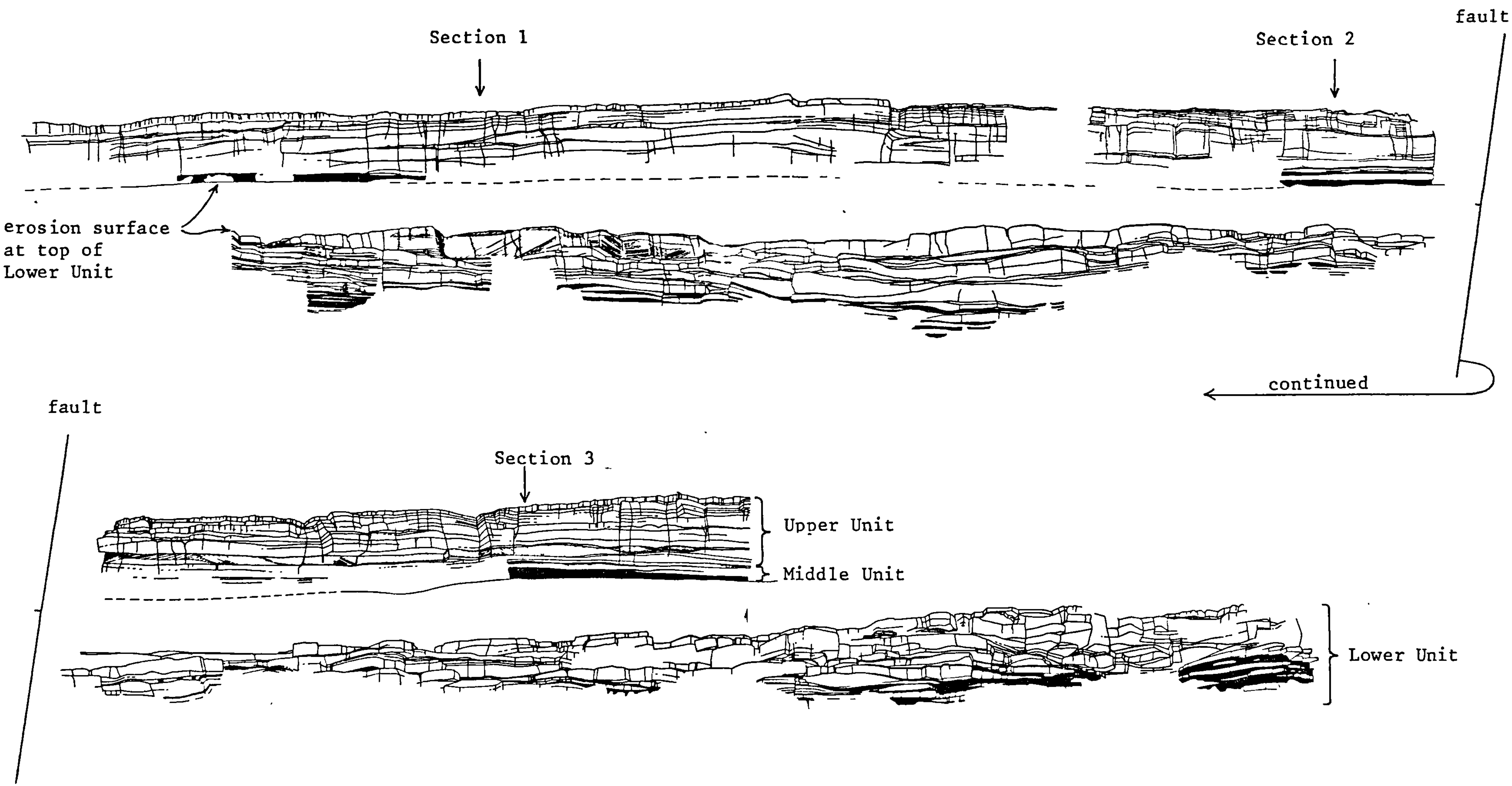


Fig.32 Sketch from photographs of the Eastern Cliff exposures of the Helaeth Sandstone (see also Plate 56); for numbered sections see Fig.33. Middle and Upper Units are drawn separate from the Lower Unit due to change in perspective.

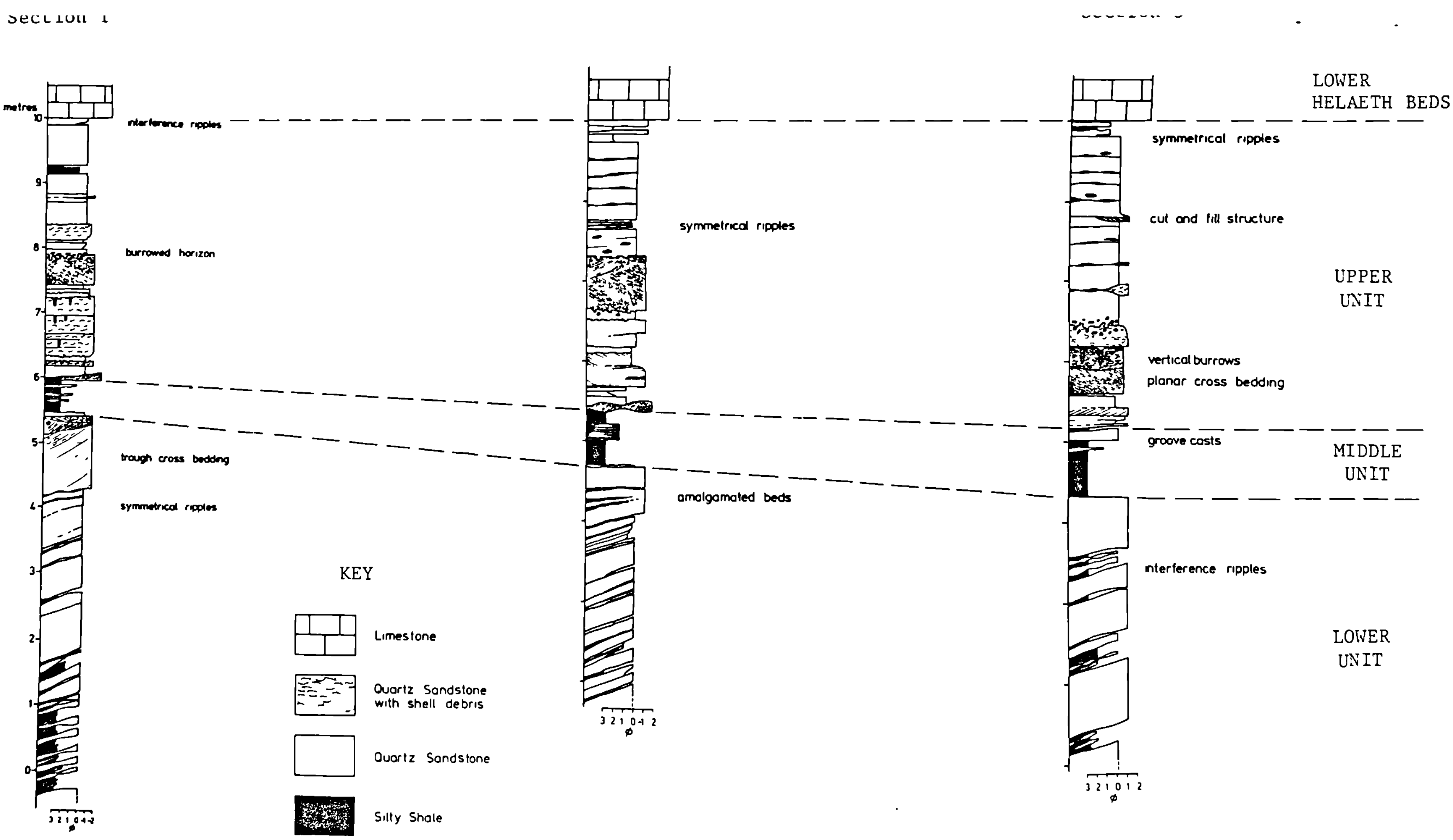


Fig.33 Graphic logs through the Eastern Cliff exposures of the Helaeth Sandstone (see Fig.32 for location of sections).

Lower Unit

The lower unit is constructed from one of the most unusual lithofacies and is unique to the Helaeth Sandstone. It consists of highly lenticular sandstone beds with a pronounced primary depositional dip to the south (Plate 59). Sections through this lithofacies are given in Fig. 33. It forms a coarsening upwards sequence. In its lower parts, shaly siltstone intercalations are thicker and the sandstones finer grained whilst uppermost sandstone beds are coarse grained with scattered quartz pebbles. These upper beds differ somewhat from the underlying lenses and are described separately.

Lenticular Sandstones : The form and dimentions of the lenticular sandstones are illustrated in Plates 59 and 60. They dip to the south at between 10^{o} and 12^{o} an attitude which stratigraphical relationships demand to be largely depositional (Plate 59b). Lenticular sandstones display intimate geometrical relationships, with swellings of many beds corresponding to troughs of underlying units (Plate 59c). They have sharp bases and whilst few appear to be erosive (thin underlying shaly siltstone bands are often preserved) several exhibit groove casts and tool marks. Larger furrows occur at the bases of some beds truncating underlying units (c.f. fig. 8 Kelling and Mullen, 1975) and sometimes display superimposed tool marks (c.f. Bridges, 1972). Elsewhere erosion is indicated by amalgamation of the sandstones and contained siltstone clasts. Larger scour features also occur (Plate 60b).

Internally, lenticular sandstones commonly exhibit a curious, form-concordant lamination which parallels the upper bounding surfaces of the beds. As the beds pinch and swell, low angle cross bedding is developed, with individual laminae, often convex upwards, abutting against the lower surfaces of the beds and either remaining conformable or asymptotic to the upper surfaces (Plate 60b). Where lenses pinch out sharply, this cross bedding is correspondingly disposed at steeper angles (Plate 61a). Such cross bedding may be directed both up and down the depositional dip, but is more often aligned perpendicular to it. No clear cut examples of more normal forset cross stratification attributable to migrating dunes or ripples and tractional currents have been recorded from these lenticular sandstones. The top 2 to 3 cms of many beds do, however, exhibit often poorly defined wave generated bidirectional cross lamination (Goldring and Bridges, 1973 and de Raaf et al, 1977) and upper surfaces of these accordingly display symmetrical ripples. Ripple crests vary from straight, with crests orientated perpendicular to the depositional dip, to complex interference patterns. Elsewhere tops are sharp and often display sinuous trails of Palaeohelminthoida (Plate 61b). Other trace fossils are rare, but the bases of a couple of beds do display well preserved examples of Chondrites (Plate 61c). No autochthonous body fossils have been recorded from these units.

Compositionally the sandstones are quartz rich with varying proportions of skeletal carbonate grains, mainly crinoidal debris. In lower parts of the unit, thinner sandstone beds are generally free of carbonate grains. Thicker and higher beds become increasingly calcareous and exhibit segregation of skeletal grains into carbonate rich bands up to 1 cm in thickness. These are weathered back preferentially to give a characteristic ribbed appearance to the sandstones (Plate 60b). Thus, whilst lenticularity, internal stratification etc. make this lithofacies very different, the sandstones are compositionally very similar to overlying calcareous sandstones of Lithofacies C.

Many of the carbonate rich sandstone units have a capping of up to 10 cms of pure quartz sandstone (Plate 59c) reminiscent of the limestone/quartzite couplets described by Kelling and Mullin (1975), though no grading of the lower calcareous parts has been observered.

Coarse Pebbly Sandstones : Calcareous pebbly sandstones form the uppermost beds of the coarsening upwards lower unit. They differ from the underlying lenticular sandstones in their coarser grain size and in the abundance of well developed trough cross-bedding (Plate 62). Transport directions to the south i.e. in the same direction as the dip of the underlying lenticular beds are indicated (Fig.30). In the far eastern part of the section these coarse sandstones comprise one bed 1.25 m thick and rest with sharp erosive contact on the lenticular beds. Towards the centre of the section, between the two faults these upper sandstones display a more gradational passage from the underlying units. They thicken and split into a broad channel feature (Plate 63) within which the basal beds display (?)antidune cross bedding (Plate 62c; c.f. Schmincke, Fisher and Waters, 1973).

In the western part of the section the coarser capping beds are absent and the dipping sandstone beds make up the whole of the lower unit (Plate 59b).

The lower unit is terminated by a conspicuous planar erosion surface (Plates 59b and 64a).

Middle Unit

This erosion surface is abruptly overlain by a thin development of shales and sharp based sandstones of Lithofacies B. These comprise the middle unit of the sequence exposed in the eastern cliff (Plate 64). The shales are limonitic and appear devoid of any body fossils. The sandstones are typical of the lithofacies with groove casts, planar lamination and primary current lineation. Tops occasionally display broad scour features filled by calcareous sandstone (Plate 64c). Palaeocurrent vectors for the unit are shown in Fig.30.

Upper Unit

The overlying upper unit of this eastern cliff section of the Helaeth Sandstone is in a sense the type locality for the calcareous sandstones of Lithofacies C. The unit exhibits to perfection the range of lithologies, sedimentary structures and trace fossils characteristic of the Lithofacies (Section 4.4c; Plates 49 and 65). Directional information provided by the various sedimentary structures in this upper unit are shown in Fig.30 and indicate a marked divergence of transport directions from those in underlying parts of the section. The unit is abruptly overlain by limestones of the Lower Helaeth Beds (Plate 59a

(iv) Environmental Interpretation for Eastern Cliffs Section

The dominantly calcareous nature of the sandstones in this section and the associated trace fossils indicate marine influence during deposition. They represent siliciclastic deposits reworked within the confines of a possibly quite broad Helaeth embayment during transgression which culminated in the deposition of the overlying Lower Helaeth Beds.

Lower Unit

Lenticular Sandstones : Any environmental interpretation for the lower lenticular sandstones must account for their primary depositional dip, but it is pertinent to discuss first the origin of the individual beds.

Sharp erosive bases, the presence of groove casts, furrows, occasional amalgamation and the occurrence of trace fossils only on bounding surfaces are thought to indicate a high energy, single event mechanism of deposition carrying coarser material into an environment where the background sedimentation was of silts settling out of suspension (Goldring and Bridges, 1973; Kelling and Mullin, 1975 and Cant, 1980). Given the clear compositional links with the beach and near-shore bar deposits of Lithofacies C and the environmental context outlined above the most likely agents to generate such units are storms (Goldring and Bridges, 1973; Kelling and Mullin, 1975; Bridges, 1976 and Vos, 1979). Sandstones attributed to a storm origin are generally latterally persistent sheet sandstones (Goldring and Bridges, 1973). The lenticular Helaeth examples clearly differ, although lenticular sandstones from other ancient shallow marine sequences have also been assigned to storm events (Kelling and Mullin, 1975 and Baldwin and Johnson, 1977).

The processes and resultant sedimentary structures which characterise storm generated sheet sands have been discussed in connection with Lithofacies B (Section 5.5a). Whilst some of these appear pertinent to an interpretation for the Helaeth units these lenticular sandstones exhibit many anomalous features. The absence of normal foreset cross stratification including current and ripple drift cross lamination and also the lack of true planar lamination as for instance compared with overlying units of Lithofacies B are thought to indicate that current driven tractional processes did not operate during sandstone deposition.

Many of the features of the lenticular sandstones may be explained with reference to hummocky cross stratified sandstones described by Harms et al (1975). These occur interbedded with silty shales and exhibit sharp, tool marked bases, internal lamination that is form concordant or 'nearly so', and wave rippled tops. More importantly they are characterised by a lack of current driven, tractional sedimentary structures. Such sandstones are attributed to strong wave action related to storm events. The internal structures of these beds is thought to result from sediment settling out of suspension under orbital storm wave motion. The wave rippled tops may form under waning storm conditions, or may represent more protracted post storm reworking. Further ancient example of hummocky cross stratification have been described by Vos (1977) and Cant (1980).

There are differences between the Helaeth beds and the described examples of hummocky cross stratified sandstone. The latter occur as tabular beds (although Cant also illustrates occasional lenticular units) and are developed in silts or fine sandstones. Perhaps the strongly lenticular form of the Helaeth units reflects the larger grain sizes involved and the correspondingly stronger storm wave action required to place such material into suspension?

Similar storm generated lenticular sandstones described by Kelling and Mullin (1975) were thought to be disposed in shallow channels compared to the storm surge channels of Brenner and Davies (1973). The scour features observed in the lower unit may well have a similar origin (see below) but the individual beds themselves although erosively based are demonstrably not channel filling displaying as they do intimate geometrical relationships between swells and troughs of succeeding beds with conformable siltstone intercalations. It is clear that extant undulating surfaces with silt veneers were inundated by ensuing units which preferentially filled the hollows and thinned over the highs.

Before going further it is necessary to consider the other feature of the Lower Unit, the depositional dip. Given the general environmental setting for these units i.e. the reworking of siliciclastic sediments during marine transgression, several types of sand body which incorporate a depositional ramp, likely to develop sequences with pronounced depositional dips, come to mind. The more obvious ones are the various delta types, fluvial and tidal. Others include various tidal or storm generated ridges developed on the near-shore shelf and also barrier complexes which may display both landward and seaward slopes. Seaward slopes of land-attached beach complexes are also possibilities. Fluvially dominated deltas are rejected on the basis of context, sediment composition, and absence of current driven tractional sedimentary structures. The latter point also dismisses tidally influenced sandbodies (Boothroyd and Hubbard, 1975).

Seaward slopes of both land attached and barrier beach complexes may produce dipping units though generally not so steep and again these would be expected to display more evidence of tractional currents and a richer marine fauna. Depositional models may be erected for both the remaining types, storm generated ridges and landward slopes of barrier complexes.

Storm generated ridges have been described from the eastern shelf of North America by Swift et al (1972) and Duane et al (1972). Helical currents generated during storm activity carry coarser sediment to the crests of the ridges, whilst the winnowed fines settle on the lower flanks and intervening troughs. Thus with lateral migration these ridges have the potential to form the coarsening upwards sequences observed in the Helaeth Sandstone. The internal structures of these modern ridges are not well documented (Johnson, 1978). Tractional sedimentary structures have been recorded but from ridges where storm generated currents predominate. Where storm wave action is as or more important incremental units of hummocky cross-bedded sand, analogous to the lenticular units in the Helaeth Sandstone, are perhaps to be expecte The tops of such units may then be reworked into small scale symmetrical ripples during waning storm conditions or by daily wave action and are subject to subsequent colonisation by burrowing benthos.

A more controversial model is required to generate the observed sequence in a back barrier setting, although the restricted ichnofauna is consistent with such an environment. Barrier complexes under the influence of marine transgression generally migrate by a process of storm washover transferring sediment from the seaward to the landward lagoon (Swift et al, 1971). A coarsening upward sequence is developed (Bridges, 1976).

A storm washover origin for the lenticular sandstone beds must, therefore, be considered. These units differ however from the sheet sandstones including those in Lithofacies B, which are normally assigned a washover origin. And are unusual not only in their lenticular form, but in the absence of current driven tractional sedimentary features. Here the work of Schwartz (1975) on modern barrier washover deposits off the east coast of the U.S.A. is of relevance. Subaerial portions of these washover fans are composed of horizontal and low angle cross stratified sands. The stratification is produced by successive storm waves topping the barrier and depositing their entrained sediment load on back barrier flats. The textural and structural properties of the sediments are explained in terms of deposition from a sediment gravity flow. Schwartz discusses this mechanism in detail and likens it to a grain flow process (Middleton and Hampton, 1973 and Sanders, 1965), a term generally reserved for the realm of deep sea fans. Grain flow deposits have the attractive property of being typically devoid of internal tractional sedimentary structures (Stauffer, 1973). According to Schwartz washover grain flow processes result in a segregation of grains into compositionally distinct laminae similar to the quartz rich/carbonate rich alternations in the Helaeth units.

These properties and effects are confined to the subaerial portions of Schwart's washover fans. Where the flow enters back barrier ponds or lagoons tractional processes start to operate and the washover fan builds out as a delta feature with internal forset cross-bedding. A similar lateral transition has been described by Warme (1971) for washover fans on the Californian coast. Schwartz argues that subaqueous washover deposits are characterised by tractional sedimentary structures and that grain flow processes do not operate. Clearly in the Helaeth Sandstone the lower unit is a demonstrably subaqueous deposit. The siltstone intercalations indicate the steady rain of suspended sediment between periods of sandstone deposition. There is evidence of reworking by a marine ichnofauna. There are no features indicative of emergence. Must then a grain flow mechanism for the emplacement of the lenticular sandstones necessarily be abandoned?

The generating mechanism of Schwart's grain flow washover events were storm wave surges, rather than the traditional mechanism of movement under gravity. Slopes of 18^{o} or higher are required for gravity induced grain flows to occur (Middleton, 1970). The lenticular beds of the Helaeth Sandstone dip at around 12^{o} and would not be steep enough to initiate grain flows. It is possible, however, that flows generated by some other mechanism could be sustained, for short distances at least, on such shallower slopes. A tentative model for the lower unit of the Helaeth Sandstone is proposed in which both the above generating mechanisms play a part.

A barrier bar, fully submerged under storm conditions is envisaged. Washover events, which breach the bar crest, are thought to provide the initial impetus or trigger for gravity driven grain flows on the landward slopes (Fig.34). Certainly one major attraction of a grain flow mode of emplacement is that it would explain the "packing" arrangement of the lenticular sandstones (e.g. Plate 60a) a feature difficult to account for by any other mechanism. Grain flows moving under gravity preferentially follow depressions in the surface over which they flow and not only fill these but tend to reverse the underlying topography forming ridges where there were depressions and depressions where there were ridges. This property is superbly illustrated by artificial grain flow deposits observed by Carter (1975).

No described modern analogues for the above model are known to the author. Most research on washover deposits deal with those emplaced onto readily accessible back barrier flats or shallow lagoons (Andrews, 1970; Schwartz, 1975 and McGowen and Scott, 1976). The steepness of the landward slopes envisaged for the Helaeth Sandstone barrier and the thickness of the lenticular deposits implies a relatively deep (10 m or more) back barrier lagoon or estuary. Such environments are not well documented. Barrier complexes off the Trucial Coast of the Persian Gulf protect comparably deep lagoons (Purser and Evans, 1973). The steeply dipping landward slopes of these barriers are developed in carbonate sand deposited by both storm and tidal activity which has accumulated close to its angle of repose (op. cit. fig.6). The internal geometry of these deposits is not well known, but they would appear an ideal candidate to examine for back barrier storm induced, grain flow sedimentation.

Coarse Pebbly Sandstones : These coarser grained, trough cross bedded units which cap the lower unit are interpreted as the barrier-bar or storm ridge crest facies. Here tractional processes and

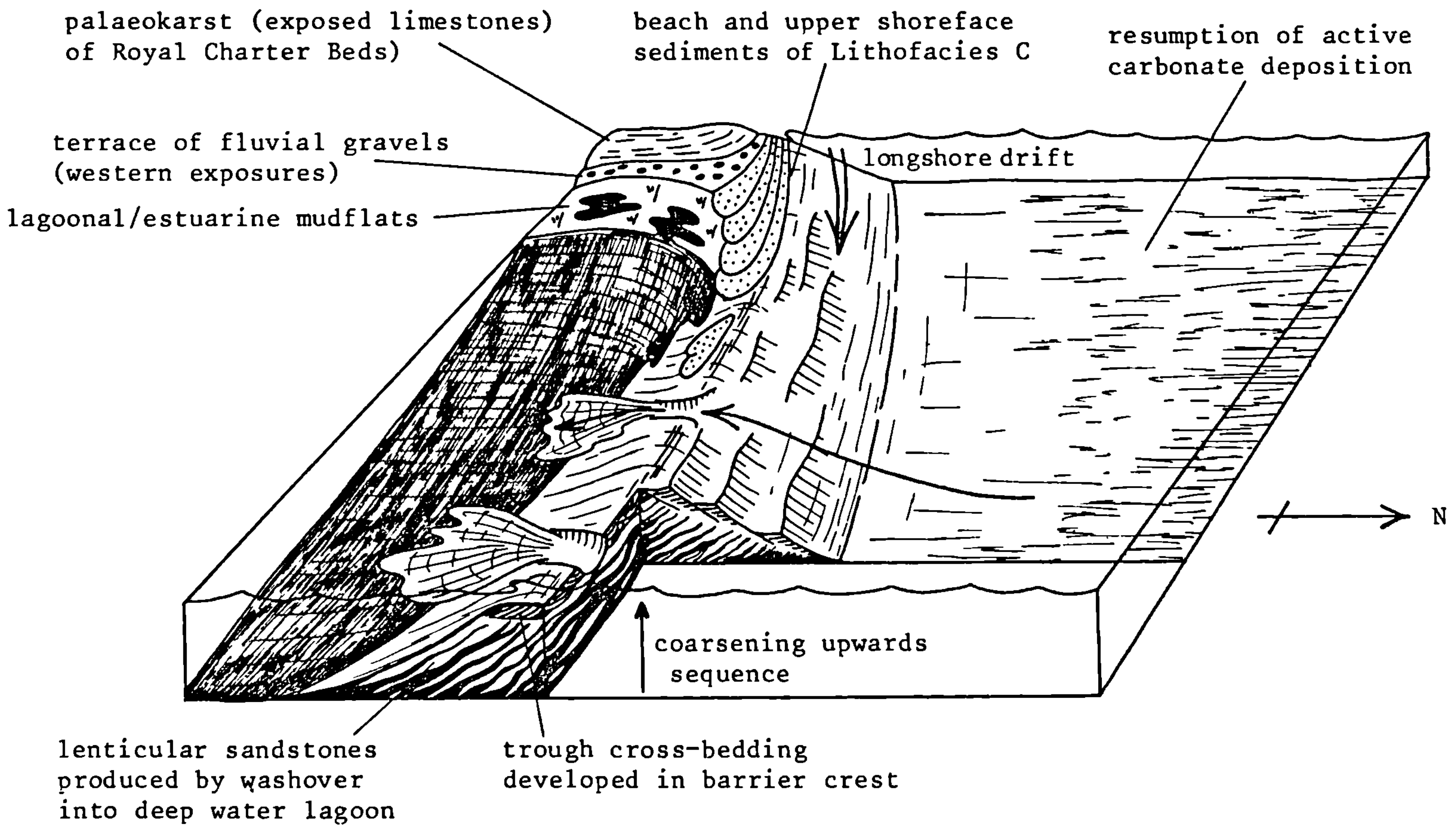


Fig.34 Suggested facies relationships during the deposition of the calcareous sandstones of the eastern cliff section of the Helaeth Sandstone.

bed forms produced by both storm and daily wave and tidal conditions predominated (Fig.34).

The channel feature towards the centre of the section into which these units thicken (Plate 63) is interpreted as a storm surge channel through the barrier crest (cf. Brenner and Davies, 1973 and fig.9.42 of Johnson, 1978). The occurrence of (?)antidune cross bedding, blocks of limestone and rolled coral colonies within the fill suggest extremely high energy conditions related to major storm or hurricane events. The scour features observed in the lower lenticular units (Plate 60b) may represent the distal parts of such storm surge channels.

The planar erosion surface which defines the top of the lower unit may represent a surf zone truncation surface (cf. Bridges, 1976) or a wind deflation surface.

Middle Unit

The overlying middle unit of Lithofacies B represent deposition under estuarine conditions (cf. Section 4.5a). The absence of any shelly fauna in the shales suggests adverse salinity conditions and indicates the re-establishment of an off-shore barrier complex. The sheet geometry of the sandstone beds contrasts with the underlying lenticular units, but in keeping with the discussion in Section 4.5 these may also be assigned a washover origin. These sheet sandstones are thought to reflect deposition on shallow, locally emergent estuarine flats in contrast to the deeper lagoonal setting discussed above.

Upper Unit

The upper unit of the Helaeth Sandstone in these eastern cliffs display in detail the range and sedimentary structures which characterise beach and upper shoreface deposits (e.g. Clifton et al, 1971

Clifton, 1973; Davidson-Arnott and Greenwood, 1976). They resemble particularly closely the suit of structures described by Clifton et al (1971) from high energy, non-barred, beaches in California, compare for instance their fig.18 with Plate 65.

The latter authors recognise a series of beach and nearshore 'facies' characterised by their hydraulic regimes and subsequent bed forms. The facies in their simplest form parallel the shore, but are often highly modified due to the effects of longshore currents. Seaward 'facies' are characterised by assymmetric ripples and larger lunate megaripples and generate landward dipping cross-lamination and trough cross-bedding. These 'facies' give way to landward swash zones and seaward migrating beach ridges, the former typified by planar lamination, the latter by seaward dipping, often low-angle and wedge shaped, cross-bedding. Lateral migration of the various facies in response to daily tidal effects, but also, presumably, with progradation or shoreface retreat (Swift, 1968) provides for an intercalation of these various sedimentary structures and the formation of complex sequences.

The combination of both landward and seaward dipping cross-stratification should result in bipolar current vectors, whilst Clifton et al also illustrate well developed 'herring-bone' cross-bedding. The latter structure is not, as normally interpreted, a function of reversing tidal currents, but due to the seaward migration of beach ridges over cross-bedded units deposited by earlier landward facing lunate megaripples. The unimodal cross-bedding directions recorded from the upper unit of the Helaeth Sandstone contrasts with these Recent findings. This may reflect the predominance of either on or offshore transport, but the divergence of the measured transport directions from those in the underlying units may indicate the operation and modifying effect of a strong longshore component.

The stratigraphic context of these upper units of Lithofacies C suggests deposition under a transgressive regime. They are thought to record the transgression of barrier beaches behind which the estuarine deposits of Lithofacies B (Middle Unit) were accumulating and from which the washover sheet-sands may have been derived (cf. Bridges, 1976). The abrupt transition into the overlying limestones of the Lower Helaeth Beds is suggestive of surf zone truncation or at least rapid 'kick-back' transgression (Irwin, 1965) and drowning of the barrier complex (Swift, 1974; Field and Duane, 1976). The contact marks the end of siliciclastic sedimentation in the Helaeth embayment and a return to active carbonate deposition on the Dinantian shelf.

(c) Other Channel Sandstones

(i) Pedolau Sandstone

See Plate 66.

(ii) Aber Sandstone

See Plates 50 and 51.

4.6 SHEET SANDSTONES

(a) Description

In the Principal Area only the Traeth Bychan (Plate 66) and possibly the St. David's and Castell Mawr Sandstones (Fig.7; Plate 15) fall into this category. These are sandstones which, whilst immediately overlying palaeokarstic surfaces, appear not to be restricted to depressions or channels and display a sheet-like geometry. They are abruptly overlain by succeeding limestone sequences and vary in thickness from 0.10 m to 2.40 m. Their lateral continuity makes them ideal marker horizons.

The distribution and contact relationships of siliciclastic units in the Straitside Area suggest that many of these are stratiform intercalations within the limestone sequence. The limitations of exposure however prevent their true geometry from being assessed. No sheet sandstones have been recognised in the Penmon Area.

Sheet Sandstones are composed exclusively of units of Lithofacies C, although anomalous trace fossils e.g. Zoophycus have been recorded from this lithofacies in these sandbodies. Where beds of Lithofacies C forming parts of channel-fill sequences extend beyond the margins of the channels onto adjacent palaeokarstic surfaces e.g. Benllech Sandstone, their contact relationships are identical to those of Sheet Sandstones (Fig.25).

(b) Environmental Interpretation

Within Channel Sandstones,units of Lithofacies C are interpreted as transgressive barrier beach and uppershore face deposits. The obvious similarities between such deposits, where they extend beyond the confines of the channels, and the Sheet Sandstones leads to an interpretation for these latter units also as transgressive sand sheets. They are interpreted as the residual deposits of siliciclastic barriers, bars and beaches which migrated across the shelf during transgression, the product of erosional shoreface retreat (Swift, 1968, 1974; Field and Duane, 1976). Anomalous trace fossils evidence later reworking of these deposits under deeper shelf conditions after drowning.

4.7 BASAL SANDSTONES

As the Lower Carboniferous succession of the Principal Area onlaps south-westwards over the older formations of the Island, a major siliciclastic unit, the Lligwy Sandstone, forms a basal deposit (Figs.6 and 8). The Lliwgy Sandstone in its strictest sense is not exposed on the north-east coast and examination is restricted to limited inland exposure. An extension of the basal deposits, however, the Lligwy Bay Sandstone (see (b) below) as its name suggests, outcrops at Lligwy Bay and forms the lowest exposed sandstone unit of the coastal section. The higher Forllwyd sandstone exhibits the same stratigraphic relationships (Fig.11).

Beyond Llangefni a series of small unconnected sandstone bodies occupy basal settings the Pencraig, Llanfawr and Henblas Sandstones.

In the Straitside Area the Fanogle Sandstone is basal in character and is equivalent to the Basement Series of the mainland side (Greenly, 1928). These are now collectively termed the Bridges Sandstone Formation but space precludes further detailed description of this unit. There is no exposed equivalent in the Penmon Area, but see Chapter 2, Section 4a.

(a) Lligwy Sandstone Formation and other basal sandbodies of the Principal Area

(i) Description

The Lligwy Sandstone forms a basal siliciclastic wedge to the Dinantian succession of the Principal Area, separating the main cyclic limestone sequence from older highly deformed basement rocks. Its outcrop extends from Lligwy Bay where it attains a thickness of around 45 m (the base is faulted against the Old Red Sandstone here and thicknesses can only be estimated) thinning south-westwards before disappearing at Llangefni. The Sandstone is not exposed at the coast and can only be examined at isolated inland exposures.

Stratigraphic aspects of the deposit have been discussed in Chapter 3, Section 4. Briefly, successive and onlapping minor cycles of the main limestone sequence can be traced towards the basal siliciclastic wedge and in this sense the junction between the Lligwy Sandstone and the limestone sequence is a diachronous one. The contact is nowhere exposed, nor is the unconformity at the base of the Formation.

Exposures in the Lligwy Sandstone are concentrated towards the north-eastern end of the crop particularly around Penrhos, Lligwy. The lowest exposures in the formation occur to the south-west of Pwll-coch [4852 8645]. Two to three metres of typical poorly sorted Lithofacies A conglomerate are observed. Pebbles with long axes up to 13 cms are present. Cross bedding and pebble imbrication are poorly developed. Further isolated exposures e.g. at Pont y Felin [4942 8634], east of Pen-y-bont [4860 8610], and south of Cae-brych [4875 8575], are all developed in conglomerates and coarse pebbly sandstones of Lithofacies A. The only significant section in the formation to show any lithological variation is provided by the quarry at Pen-y-bont [4810 8603]. Fig.38 illustrates the sequence observed in the southern portion of the quarry. Massive conglomerates predominate and are conspicuous for the coarseness of the plant debris they contain (Plate 44c). Finer sandstone lithologies are present and allow the recognition of at least two fining upwards units. In the northern half of the quarry, on the downthrown side of a small fault, a clay seam intercalated with the normal conglomerates is observed. The clay varies from black and carbonaceous, with abundant small (< 1 cm) pyrite nodules, to red with pale grey haloes around branching carbonaceous filaments, the latter interpreted as fossil rootlets.

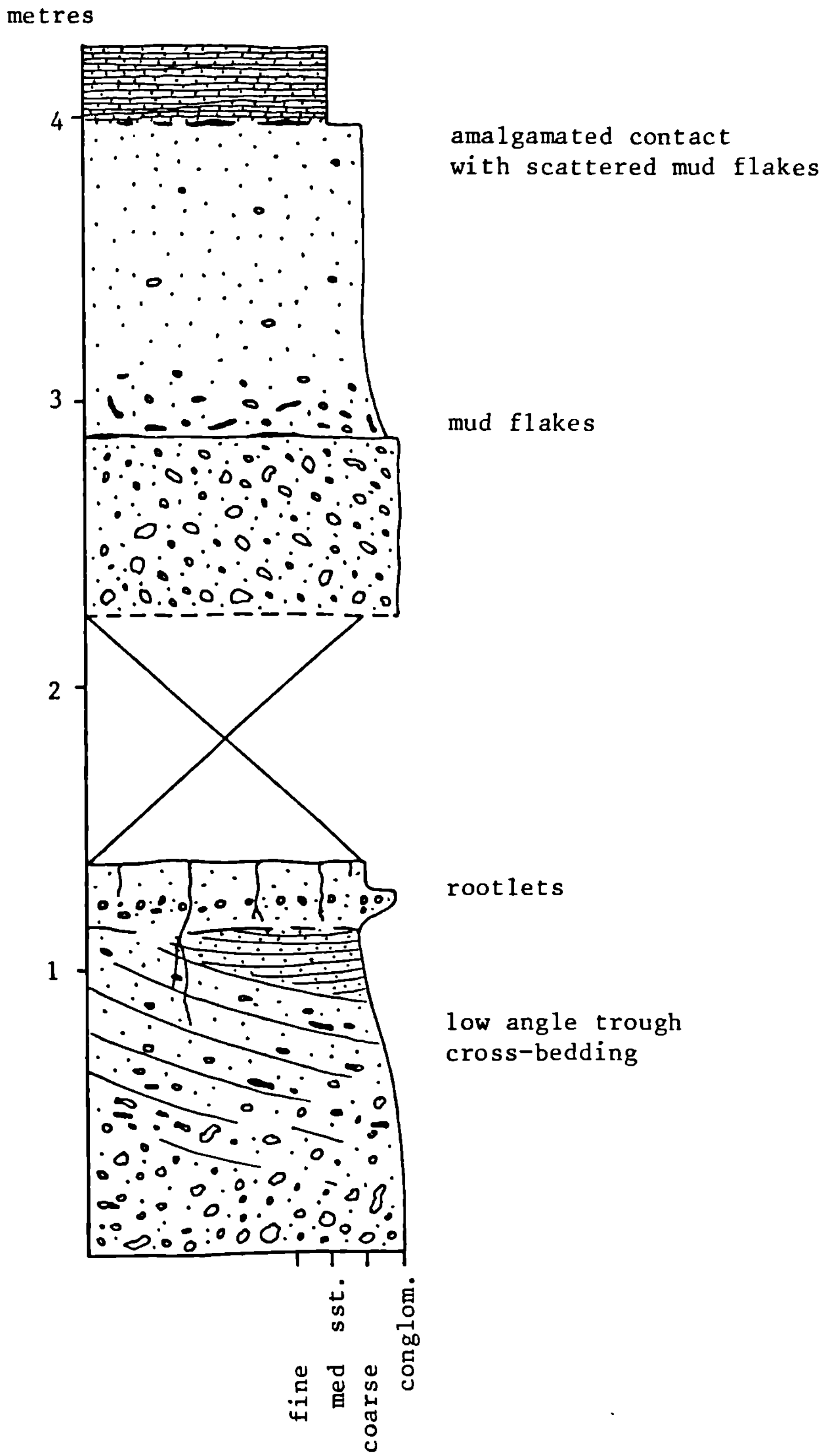


Fig.38 Small scale fining upwards sequences in the Lligwy Sandstone, Pen-y-bont Quarry [4810 8603].

Towards the south-west, the crop of the Lligwy Sandstone is marked by low marshland which has recently been extensively forested. Lithofacies A conglomerates are exposed to the east of Tyddyn-isaf [4833 8530] and finally to the south of Capel Babell [4745 8335]. South to Llangefni there are no further natural exposures. Temporary sections close to the point where the Lligwy Sandstone pinches out were provided in house foundation trenches, to the north of Llangefni, at Pont Clai [4625 7622]. These revealed up to 3 m of coarse breccia deposits composed of large angular blocks of Precambrian quartzite and highly decomposed and reddened chlorite schist. These latter units appeared to be abruptly overlain by 3 to 4 m of thinly bedded to nodular limestones with shale bands, locally rich in Lithostrotion junceum and referred to the Traeth Bychan Limestone Formation.

The few palaeocurrent vectors obtained from the Lligwy Sandstone Formation indicate transport paths to the east and north-east.

From Llangefni to the south-west a series of isolated sandstone bodies occupy a basal setting (Fig.8).

The distribution of the Pencraig Sandstone suggests that at depth it is probably contiguous with the Lligwy Sandstone outcropping to the north, but at surface forms a distinct and separate unit. The quartzite inlier it surrounds demonstrates the irregular nature of the sub-Carboniferous floor. Mapped relationships indicate tongues of Pencraig conglomerate penetrating the local limestone sequence at a level thought to correspond to the palaeokarstic surface separating the Upper and Lower Dinas Beds, the horizon of the Benllech Sandstone at the coast. There are few extant exposures in the unit. Coarse, poorly sorted, massive conglomerates are observed in disused and largely filled quarries at [4657 7532] and [4663 7570]. Large quartzite boulders in the latter locality reflecting the proximity of the Precambrian inlier.

The Llanfawr Sandstone recognised by Greenly (1919) is no longer exposed, its outcrop pattern is based largely on features.

The Henblas Sandstone is intermittently exposed in the stream of that name [4320 7223]. The deposit consists of very coarse conglomerates with angular to sub-rounded blocks (up to 20 cm diameter) of Mona Complex lithologies (quartzite, schist etc.) set in a matrix of fine, orange-brown ferruginous sand. Intercalated with the coarser deposits are friable, mottled green and purple chlorite flake breccias derived from decomposed and disaggregated blocks of the local chlorite schist.

(ii) Environmental Interpretation

It is clear from the above description that there is little information on the sequential variation of lithologies within the Lligwy Sandstone to allow detailed environmental analysis. Critical exposures at the base of the formation and more importantly, showing the passage into the main limestone sequence are also absent. An assessment of the nature of these junctions rests with mapped relationships and other more circumstantial stratigraphic evidence (see below).

The predominance of conglomerates and pebbly sandstones of Lithofacies A in comparison with such units in the Channel Sandstones (Section 4,5a) suggests that the Lligwy Sandstone is largely an accumulation of fluvially deposited sediment. Moreover the coarse grainsize and the paucity of fines is again indicative of braided fluvial systems. The fining upwards units noted from Pen-y-bont quarry resemble those described by Smith (1970) and Vos (1977) from ancient siliciclastic sequences also assigned a braided river origin and thought to reflect the fluctuating discharge which characterises such fluvial regimes. The clay seam observed at the same locality may represent the rarely preserved muddy and vegetated fill of an abandoned braid channel (Williams and Rust, 1969).

Combined with the gross stratigraphic context of the basal sandbody the presence of braided stream deposits leads to an intuitive model for the Lligwy Sandstone as an alluvial fan complex. Siliciclastic debris derived from the adjacent hinterland of older rocks are envisaged accumulating along this north-western margin of the Dinantian cuvette as a series of coalescing alluvial fans or cones whilst the abundance of coarse plant debris observed within the formation (Plate 44a) suggests that these fans were in parts at least extensively vegetated. Depositional models errected for the Van Horn Sandstone by McGowen and Groat (1971) are applicable though probably on a much smaller scale. Their proximal fan deposits, boulder conglomerates, are not recognised in the Lligwy Sandstone although the breccias of angular quartzite and schist blocks observed at Llangefni probably represent scree deposits accumulating in a proximal setting. The bulk of the exposed Lligwy Sandstone appears to fall into McGowen and Groat's mid and distal fan facies where the processes and deposits of braided streams were thought to predominate.

Critical to our better understanding of the Lligwy Sandstone and of siliciclastic deposits within the Anglesey Dinantian in general is the nature of the transition between the basal sandbodies and the main limestone sequence. Successive formations and their constituent minor cycles can be traced towards the basal sandbody

(Figs. 8 and 11) and in this sense the contact is a diachronous one. Again intuitive models come to mind, but an insight into the actual lithofacies relationships within this transitional zone is provided by the Lligwy Bay Sandstone, an extension of the main basal clastic wedge and it is therefore appropriate to discuss this unit first.

(b) Lligwy Bay Sandstone

(i) Description

Sandstone exposures at Lligwy Bay including the famous conglomerate of the same name (Greenly, 1919) have hitherto been regarded as the uppermost unit of the basal Lligwy Sandstone (Greenly, 1919; Mitchell, 1964; George , 1975). Natural movement of beach sands has recently extended the outcrop here and shown the Lligwy Bay Conglomerate and associated sandstones to rest on a deeply eroded limestone platform (Cope, 1975). It is, therefore, arguable that these siliciclastic deposits represent a separate and higher unit than the Lligwy Sandstone proper. The unit is therefore termed the Lligwy Bay Sandstone. Discrete though this unit may be at the coast, it must, when traced inland, rapidly merge with the basal deposits. Fig.11 demonstrates the probable stratigraphic relationships in the area and prompts an interpretation for both the Lligwy Bay Sandstone and the higher Forllwyd Sandstone as distal extensions of the Lligwy Sandstone rather than more discrete channel fills. Implicit in such an interpretation is the uninterrupted passage of lithofacies from these sandstone units into the basal deposits where as lithofacies development within the channels was of a more independent and isolated nature. The vertical sequence at Lligwy Bay provides an insight into the lithofacies distribution within the true basal sandstones where they come into contact with the adjacent limestone succession, a transition otherwise unexposed.

The section begins at Careg-ddafad in the southern corner of Lligwy Bay where the outcrop of the spectacular Lligwy Bay Conglomerate has been described by Greenly (1919, p.602 and 629), Mitchell (1964) and Cope (1975). This unique deposit which attains a thickness of approximately 2 m is now recognised as the basal unit to the Lligwy Bay Sandstone and rests on an irregular limestone surface. This surface exhibits characteristic evidence of emersion (Chapter 3), it is of high relief with a conspicuous upstanding block of limestone 1.5 m high and 3 x 4 m in plan around which the conglomerate is banked (Plate 67a and fig.6 of Cope, 1975). The Lligwy Bay Conglomerate is an exceptionally coarse grained and poorly sorted deposit aptly termed by Greenly a "boulder bed" with clast sizes ranging up to 60 cms across and with a coarse sandstone matrix. The composition and provenance of the pebbles and boulders are discussed by Greenly (1919, p.567 and 602). Some of the largest are angular blocks of limestone, often highly dolomitised, of local derivation. The rest are made up of Ordovician grits and mudstones and various pyritised and silicified rocks which Greenly thought derived from the metasomatic tract of Parys Mountain, 6.5 kms to the north-west. Pebbles of these latter lithologies are generally well rounded. The absence of clasts derived from the Mona Complex is conspicuous and with the coarseness of the grain size sets the deposit apart from more normal conglomerates of Lithofacies A which overlie it. At the top of the boulder bed and within these overlying units are well defined lenses of coarse sandstone up to 6 m across and 35 cm thick

(Plate 67b). Internally these display a crude horizontal to low angle inclined stratification and their tops exhibit pebble filled scours. Similar features have been described by Howell and Link (1979) from Eocene conglomerates in California.

Beyond Careg-ddafad towards the cliffs on the south side of Lligwy Bay the section becomes increasingly complicated with carbonaceous shales pinching out against large irregular masses of conglomerate (Plate 67c). This zone of complexity culminates in the extraordinary Lligway Bay Disturbance of Greenly (1919, p.615) which occupies the western end of the main cliff section and which is thought to demonstrate the effects of solution and collapse within the succession (Section 3.5; Plates 40 and 41).

Past these disturbances the steep cliffs extending to the north-east display a more regularly bedded sequence (Fig.11). This is assumed to overlie the conglomeratic units seen at Careg-ddafad although, with the intervening complexities, exact relationships are necessarily uncertain. At the foot of the cliff, sparingly exposed amongst the boulders, are coarse grained sandstone referable to Lithofacies A. These are overlain by a thick unit of sulphurous shales with thin massive and cross laminated siltstone and fine micaceous sandstone beds and containing plant debris and the trace fossil *Zoophycus*. These units are assigned to Lithofacies B. Calcareous sandstones of Lithofacies C follow abruptly, they display the usual vertical escape traces and "U" shaped burrows and exhibit characteristic honeycombe weathering (Plate 48c). Sedimentary structures are not common though faint traces of cross bedding partially obliterated by bioturbation do occur (Plate 48c). Lenses of coarser shell debris are also present. With decreasing sand content these sandstones grade into overlying limestones of the

Lligwy Beds, the lowest of the limestone cycles of the Principal Area exposed on the north-east coast. Significantly as the passage beds between the sandstones and limestones are traced seawards i.e. away from the crop of the Lligwy Sandstone, a reduction in sand content also occurs.

(ii) Environmental Interpretation

The internal distribution of lithofacies within the Lligwy Bay Sandstone is not dissimilar to that observed within the Channel Sandstones i.e. basal conglomeratic units of Lithofacies A, overlying beds of Lithofacies B and capping calcareous sandstones of Lithofacies C. Yet the mapped relationships of the Lligwy Bay Sandstone and indeed of the higher Forllwyd Sandstone (Fig.11) clearly set these units apart from the Channel Sandstones encountered higher in the succession.

Despite the differences in stratigraphic setting there seems little reason to change the environmental interpretations established for the various lithofacies where they form part of channel filling sequences. Worthy of more particular attention, however, is the Lligwy Bay Conglomerate at Careg-ddafad, since this differs from the overlying 'normal' conglomerates of Lithofacies A. The abundance of clasts derived from Parys Mountain combined with the large grain sizes suggested to Greenly (1919, p.627) that this "tumultuous accumulation" was the product of torrential rivers and that deposition "would certainly have been performed in quite a short time". Certainly the provenance directions of the clasts i.e. from the north-west, differ from those of the Mona dominated 'normal' conglomerates and contrast with palaeocurrent data for the Lligwy Sandstone, both the latter indicating derivation from the west and south-west. Such evidence indicates the temporary breaching of some intervening watershed. This fluvial model contrasts with that put forward by Cope (1975) who interpreted the boulder bed as "a massive beach deposit". The rounding and sorting characteristics of the unit are, however, more in keeping with a fluvial origin. Greenly also points out that the pyrite-rich boulders derived from Pary's Mountain would quickly weather and that their state of preservation required rapid deposition. This again is suggestive of fluvial processes rather than protracted reworking within coastal littoral zones.

The overlying 'normal' conglomerates of Lithofacies A in comparison with like units in Channel Sandstones are thought to indicate the establishment of more permanent braid plain conditions with clasts derived from contemporary tracts of Precambrian strata exposed to the west. The lenses of coarse sandstone within these conglomeratic units are compared with similar features recorded from modern braided river gravels (compare Plate 67b with fig.20 of Williams and Rust, 1969) and are thought to represent sandy bar forms or channel fills (Howell and Link, 1979).

Units of Lithofacies B observed within the Lligwy Bay Sandstone characteristically reflect waning terrestrial influence and the establishment of at least quasi-marine conditions as indicated by the occurrence of Zoophycus. Overlying units of Lithofacies C compare closely with their counterparts in Channel Sandstones e.g. Helaeth Sandstone, the suite of sedimentary structures and trace fossils indicating a similar beach and upper shoreface environment of deposition (Section 4.5a). In contrast with channel filling sequences the Lligwy Bay Sandstone exhibits intercalation and gradational contacts with overlying limestone strata suggesting only the gradual establishment of carbonate shelf conditions following transgression, and indeed the interdigitation of such environments with nearshore zones of sustained terrigenoclastic deposition.

4.8 SILICICLASTIC DEPOSITION DURING THE ANGLESEY DINANTIAN

As an extension of the Lligwy Sandstone the Lligwy Bay Sandstone, and the lithofacies relationships it displays, provide an insight into the otherwise unexposed transition between the basal sandbodies and the main limestone sequence. The evidence is suggestive rather than conclusive, but the Lligwy Bay Sandstone is the final piece in a jigsaw which allows the more precise understanding of siliciclastic deposition within the Anglesey Dinantian particularly in the Principal Area.

During the many regressive episodes which punctuate the Anglesey Dinantian sequence (Chapters 2 and 3) the falls in erosive base level led to rejuvination of terrigenoclastic source areas within the hinterland of older rocks. This increased supply of siliclastic sediment was shed first onto the basal sandstones alluvial fan complexes envisaged lining the margins of the Dinantian cuvette. It seems likely that during such periods the fans would have actively prograded onto the adjacent limestone shelf forming lobe-like extensions comparable with the Lligwy Bay and Forllwyd Sandstones. Streams flowing off these fans incised into the newly lithified carbonates and established fairly permanent anastomosing channel systems (Section 3.2c). It is within the confines of these channels that transportation and deposition of sediment carried beyond the alluvial fans largely took place (Fig.10).

Marine transgressions caused first the drowning of these channel systems and the establishment of estuarine conditions as evidenced by units of Lithofacies B. As transgressions advanced barrier beaches and spits of Lithofacies C were driven across the estuarine flats forming capping units to the Channel Sandstones.

Transgressive rises in sea level culminated in the repeated inundation of the limestone platform, raised base levels, caused a concomitant reduction of terrigenoclastic supply and allowed carbonate deposition always to predominate during these periods.

Here the exposures at Lligwy Bay are critical. The capping units of Lithofacies C suggest that during these periods of raised sea level the toe deposits of the marginal alluvial fans were reworked into nearshore bars and beaches giving way rapidly seawards to areas of active carbonate deposition. Any siliciclastic sediment still shed off the fans was entrained within these nearshore sandbodies and terrigeno-clastic contamination of the carbonate shelf was thus reduced. In this context the baffling effects of coastal swamps may also have been critical preventing all but the finer suspended fractions from escaping to the outer shelf. The carbonaceous rich units of Lithofacies B observed in the Lligwy Bay Sandstone may represent such quasi-marine swamp deposits with perhaps the calcareous sands of Lithofacies C forming protective barrier beaches and spits.

CHAPTER FIVE

FACIES ANALYSIS OF THE LIMESTONE SEQUENCE

5.1 INTRODUCTION

In providing both rigid stratigraphic and environmental constraints the minor cycles, defined by palaeokarstic surfaces, provide an ideal framework for the facies analysis of the limestones. Previous studies (Greenly, 1919; Nichols, 1962; Mitchell, 1964) lacked such constraints without which the distribution of limestone types often appears random. Various carbonate lithologies were described and interpreted but in isolation without reference to their sequential arrangement. Moreover these studies came at a time when the main emphasis of limestone petrography was directed towards diagenesis, a trend inspired by Bathurst's early work on these very strata in North Wales (1958, 1959). Environmental interpretation by these previous researches is restricted to general comments on depth of water, salinity and climate.

The present study has taken advantage of the great volume of recent literature detailing the sediments and processes of modern carbonate depositing environments. This has allowed the more precise description and interpretation of the various carbonate lithofacies which make up the Anglesey succession whilst an appreciation of their sequential relationships has permitted close comparison with recent analogues.

.2 LIMESTONE CLASSIFICATION

During the tenure of this research several carbonate classifications have been tried including those of Grabau (1904), Folk (1962) and Dunham (1962). Of these the latter textural based scheme of Dunham has been found the most rewarding. The distinction between grain and matrix support is one that can be readily applied in the field and combined with an assessment of the presence or absence of mud grade carbonate has allowed the construction of graphic logs (Charts 1 to 9). The textural sequences thus portrayed are comparable to fining and coarsening upwards sequences of siliciclastic schemes (e.g. Selley, 1970) which record, in their simplest sense, variations in depositional current regimes. In limestones the largely autochthonous origin of the constituent grains means that analogous textural variations reflect more precisely changes in the potency of "currents of removal" (Dunham, p.111), rather than of supply although higher energy carbonate deposits e.g. cross-bedded grainstones are more closely comparable with their siliciclastic equivalents. Greater flexibility has been achieved by the use of dual naming system e.g. wackestone/packstone and packstone/grainstone, a reflection of the, at times, uncertainty of field identification but also of the often patchy distribution of textural types. For detailed discussion of the merits and drawbacks of various carbonate classifications including Dunhams see Ham and Pray (1962) and Flugel (1982).

5.3 GRAIN TYPES

In addition to the basic textural categories of the limestones more detailed subdivision is achieved on the basis of the nature and abundance of the constituent grains or allochems (Folk, 1962). In most carbonate classifications the same relatively few grain types are recognised (see A.A.P.G. Memoir 1, 1962; Flugel, 1982). Most are identifiable in the field and include skeletal grains (or bioclasts), peloids, ooids, oncoids and intraclasts. Peel and thin section examination of the Anglesey limestones has allowed the differentiation of the various skeletal grains into contributory taxa, and has further allowed the recognitition of grapestone-like aggregates.

(a) Skeletal Grains

Such grains include both whole and fragmental valves and tests of calcareous shelly fauna. Grains of comminuted macrofauna are derived principally from brachiopods and echinoderms whilst certain lithofacies are rich in dasycladacean algal debris. Molluscs, corals, bryozoa and trilobites all contribute to skeletal grain assemblages. For distinguishing criteria see Horowitz and Potter (1971), Bathurst (1975), Flugel (1982). Microfaunal elements include the virtually ubiquitous whole tests of foraminifera whilst in some lithologies calcispheres and valves of ostrocods predominate.

The pseudo-punctate structure of most brachiopod fragments and the frequent occurrence of spines, circular to eliptical in thin section, demonstrate derivation from spinose strophomenid genera (chonetids and productids). Spiriferid brachiopods with their distinctive prismatic shell structure are also represented.

The conspicuous single calcite crystal plates of echinoderms abound in many of the lithofacies. In hand specimen most such grains are observed to be crinoid ossicles but plates and spines of Archaeocidaris have also been collected and show that tests of echinoids provided an additional source.

Grains of dasycladacean algae are derived from calcareous skeletons which formed as essentially inorganic precipitates around the original plants (Bathurst, 1975). Within Asbian strata Koninkopora was the major contributor of such grains whilst in the Brigantian Nanopora filled the same niche.

Most skeletal grains exhibit micrite envelopes to a greater or lesser extent, and evidence the activities of endolithic boring algae and fungi

(Bathurst, 1966). Such grains represent the cortoids of Flugel (1982) The recognition of skeletal grains derived from molluscan taxa (bivalves, gastropods and cephalopods) is largely dependent on the presence of micritic rinds since the original molluscan aragonite has been replaced by sparry ferroan calcite cement. Micrite envelopes preserve the original outlines of these grains. No internal structure or organic lamellae, as observed in some Mesozoic and Quaternary bivalves (e.g. Wilson, 1967), are preserved. This suggests that the transformation aragonite to calcite was not achieved by neomorphic processes (Bathurst, 1975, p.489), indeed the occurrence of collapsed micrite envelopes indicates an intervening void stage.

(b) Peloids

The term peloid is used to describe generally rounded to subrounded micritic grains of potentially diverse origin. Most reflect the effects of complete micritisation due to boring algae i.e. micrite envelopes so advanced that the character of the original host grain is indeterminate. In most instances the initial grain was probably skeletal although extensively micritised ooids gradational into peloids have also been recorded.

A second major origin for such micritic grains is as faecal pellets, generally described as well rounded and ovoid in shape and often well sorted (Flugel,1982). Suprisingly few well formed faecal pellets have been identified from the Anglesey limestones reflecting perhaps the effects of subsequent diagenesis. The assumption has always been that early submarine cementation provides such pellets with a high preservation potential. Shinn (1973) however has described Recent subtidal carbonate sediments from the Persian Gulf in which pellets are "squashed beyond recognition", and this is an area where submarine cementation forming hardgrounds is an active process (Shinn, 1969). The effects of later recrystallisation may also blur the outline of perhaps already deformed faecal pellets and therefore preclude their confident identification.

(c) Ooids

See Lithofacies IV, Section 5.5.

(d) Oncoids

Oncoids comprise subspherical to irregular particles up to 5 cms across, composed of concentric, crinkly laminae coating a nucleus, usually skeletal. They are formed by the growth and sediment binding action of blue-green algal mats around grains which are subject to periodic movement. Infrequent movement may lead to strong asymmetry of the algal coats (Flugel, 1982).

(e) Intraclasts

Intraclasts are eroded lumps of previously lithified limestone. They represent rip up clasts and are associated with high energy often cross-bedded deposits. In the Anglesey Dinantian such particles commonly occur towards the base of minor cycles where they often comprise lumps of calcretised limestone, up to several centimetres across, derived from underlying palaeokarstic horizons. Of the same derivation are large skeletal grains with adhering micritic material (calcrete). These are also classed as intraclasts, but are comparable with the lithoskels of Read (1974) described as skeletal grains "reworked from . . . sediments of an earlier sedimentary cycle". With reduced calcrete coating such particles are easily confused with primary skeletal grains.

(f) Compound Grains

Compound grains which comprise two or more usually highly micritised and well rounded grains bound together by micrite cement are closely comparable to modern grapestones (Illing, 1954). The latter abound in Recent sediments on the Bahama Bank and provide evidence of early submarine cementation (Bathurst, 1975 p.316). They differ from intraclasts in not displaying erosional truncation of the constituent grains so that the particle as a whole exhibits a characteristically botryoidal outline.

5.4 DIAGENESIS

Diagenetic aspects of the Anglesey limestones have been described by Nichols (1962) and Mitchell (1964), whilst more general studies on the Dinantian limestones in North Wales are provided by Orme and Brown (1963) and Oldershaw and Scoffin (1967); with much of Bathurst's classic work on diagenesis based on his observations from these same strata (1958, 1959 and 1975). There seems little point, therefore, in protracted discussion of already well documented phenomena, although a clear understanding of these effects is essential if the original character of the limestones is to be appreciated and their environmental analysis facilitated. The principal diagenetic effects evident within the limestones of the Anglesey Dinantian are summarised below. The special case deserving of more detailed description and interpretation are the early alteration phenomena associated with palaeokarst formation and these have already been dealt with in Section 3.4.

(a) Pressure Solution

The many and varied effects of pressure solution are evident throughout the succession, often reflecting the character of the host limestones. Recent major studies of pressure solution phenomena have been presented by Logan and Semeniuk (1976) and Wanless (1979), and those observed within the Dinantian limestones of Anglesey compare closely with the various categories described by these authors.

Pressure solution acting on argillaceous limestones leads to the rapid concentration of impurities and the formation of anastomosing shaly partings, seams and general nodularity. These effects correspond to the effects of non-sutured seam solution of Wanless and to the stylonodular structures of Logan and Semeniuk (see Lithofacies I below). More obvious pressure solution effects are displayed by cleaner limestone types in which stylolites and sutured grain contacts abound. These are formed by Wanless' sutured seam solution and include the distinctive stylobreccias of Logan and Semeniuk. The latter are particularly prevalent within the Anglesey limestones. In the field light brown clay is often concentrated along such stylolitic contacts, whilst in cut block this material is seen to be derived from the weathering of thin films of argillaceous impurities and ferroan dolomite. Wanless has shown that the latter mineral is an anticipated product of pressure solution and grows in response to the liberation of Mg^{2+} ions.

(b) Recrystallisation

The distinction between void filling cement and recrystallised fine grained matrix material is of course critical in the textural determination of limestone types. In this context the fabric criteria of Bathurst (1975, p.484) for the identification of neomorphic spar have been rigorously applied. The effects of such neomorphism are widespread within wackestone and packstone lithologies affecting both matrix and grain alike, whilst difficulties in distinguishing these effects in the field account in part for the use of the dual packstone/grainstone category. Characteristic products of neomorphic recrystallisation include the alteration of micrite to microspar, the growth of stellate masses of radial-fibrous spar (radiaxial fibrous calcite of Kendal and

Tucker, 1973, has not been recorded), syntaxial overgrowths on echinoderm grains set in micrite (or now microspar) matrix and the common occurence of structure grumeleuse of Cayeux (1957) (see Bathurst for detailed discussion of these various phenomena).

(c) Cementation

An obvious corollary of the above discussion on recrystallisation is the recognition of fabrics in calcite spar which result from void filling cementation. Again the criteria of Bathurst (1975 p.417) have been applied. In wackestone lithologies calcite cements are confined to intra-particle voids e.g. chambers in foraminifera. In packstone and packstone/grainstone textural units inter-granular cements also occur whilst in grainstone the latter, by definition, predominate.

Staining of the various cement fabrics reveals the presence of both non-ferroan and ferroan calcite, the former invariably forming thin palisade fringes around constituent grains; the latter occurring as a late stage blocky pore filling variety. Light/dark alternations in stain intensity reveal the often complex growth stages of individual cement crystals.

(d) Dolomitisation

The present study has steered away from a detailed assessment of the distribution and origin of dolomite within the Anglesey Dinantian since this constitutes a major research topic in its own right. Future work on such aspects of the succession should take account of the recent SEPM Special Publication No.28 on "Concepts and Models of Dolomitisation", edited by Zenger, Dunham and Ethington (1980).

The effects of patchy replacement by dolomites are evident through the sequence and are made conspicuous by the mineral's usually rusty weathering ferroan composition. The wholesale replacement of limestone strata in highly faulted terrain e.g. Trwyn-du in Penmon, combined with the often coarse saccaroidal texture indicate a late epigenetic origin for much of this dolomitisation. Scattered rhoms in the vicinity of stylolitic seams, as mentioned above, may be a by-product of pressure solution (Wanless, 1979).

The common occurrence of dolomitised limestones underlying palaeokarstic surfaces is suggestive of formation during the penecontemporaneous regressive episodes these surfaces mark (cf. Ramsbottom, 1973). No textural or fabric criteria to support such an early origin have been identified and saccaroidal textures still predominate. A porosity or permeability control by the palaeokarstic horizons on later dolomitising fluids is indicated, but it is perhaps also possible that this later dolomite seeded on finely disseminated syndepositional dolomite already present with the host limestone. Clearly much more work, including detailed geochemical and isotope investigation, is required before these problems may be resolved.

(e) Silicification

As with dolomitisation, the effects of silicification within the Anglesey Dinantian succession though widespread have not been assessed in detail. The principal product of silicification is chalcedonic silica occurring as scattered, often void filling blebs, as beekite replacing fossils and as chert nodules. Some of the chalcedonic silica within the succession is length slow and it has been suggested that such varieties reflect precipitation from highly alkaline solutions (Folk and Pittman, 1971). Euhedral quartz prisms are also common often cross cutting skeletal grains and indicating a replacive origin. Chert nodules become increasingly abundant in the upper parts of the sequence and culminate in the development of the bedded cherts exposed on Castell-mawr (Fig.14) and which cap the Dinantian sequence on Anglesey. These upper units correspond to similar cherty strata developed at this same stratigraphic level across the North Wales mainland (e.g. Morton, 1886). The nature and origin of silicification in these latter units has been discussed by Sargent (1923) whilst a more up to date description and discussion of silicification as it affects Dinantian limestones is given by Orme (1973).

5.5 CARBONATE LITHOFACIES

In distinguishing the various carbonate lithofacies which make up the Anglesey succession a strong emphasis has been placed on criteria recognisable in the field, i.e. gross textural and bedding characteristics. This has allowed a ready appreciation of their stratigraphic distribution and sequential and lateral relationships. Exhaustive petrographic subdivision into numerous microfacies (c.f. Sadler, 1966) is possible, but to some extent out of tune with the relatively few lithofacies which characterise modern carbonate shelves and platforms (Ginsburg and James, 1974). Such microfacies are perhaps better regarded as showing the range of textural and compositional variations present within individual lithofacies.

The following lithofacies have been recognised:

I : Argillaceous skeletal wackestone/packstones and intercalated shales

II : Skeletal packstone/grainstones

III : Skeletal pelloid grainstones

IV : Ooid skeletal grainstones

V : Calcite mudstones

In addition to these principal textural types various subfacies Ia, IIa and IIIa have also been distinguished on the basis of their different bedding characteristics or grain assemblages. Grain counts provide quantatitve information on the composition of the various lithofacies but such statistical data are of only limited value in environmental interpretation (Bathurst, 1975 p.138).

(a) Lithofacies I : Argillaceous skeletal wackestone/packstones and intercalated shales

(i) Description

The lithofacies forms an important element of minor cycles which comprise the Traeth Bychan Formation and within these it invariably comprises a basal phase up to 12 m thick (Charts 6 and 7). The lithofacies is composed of thinly, but well bedded, dark grey, foetid argillaceous wackestone/packstones with intercalated bands and locally thick beds of shale (Plate 68). Limestone beds range from 5 to 30 cms in thickness. They may be tabular and parallel sided or exhibit undulating tops and bases and display a more nodular style of bedding. In the latter case the intervening shale bands correspondingly pinch and swell, but rarely exceed 5 cms in thickness. Thicker beds of shale up to 2 m thick do occur e.g. Moryn Beds (Chart 3). The bedding characteristics of the lithofacies are closely comparable with those developed within the Lower Lias of southern England and South Wales and described by Hallam (1964) and Wobber (1967). Tops and bases to limestone beds which appear quite sharp and abrupt in the field are seen in cut blocks to in grade over 2 to 3 mm into adjacent shale bands Isolated limestone nodules feather into their host shales whilst thicker limestone beds exhibit impersistent and anastomosing shaly partings. These latter features are reminiscent of the 'horse-tail' stylolites of Roehl (1967).

Petrographic analysis of the limestones (see Fig.41) shows them to be composed of interspersed finely comminuted skeletal debris and micrite such that the depositional texture ranges between wackestone and packstone (Plate 70b). Where larger skeletal elements i.e. the tests and valves of macrofauna (see below) are present they generally float within this finer matrix material (skeletal floatstones of Embry and Klovan, 1971, or whole fossil wackestones of Wilson, 1975). Their fine size combined with the effects of subsequent micritisation makes much of the skeletal debris within the matrix indeterminate only the distinctive fragments of echinoderm are readily identified. The contained microfauna is dominantly of foraminifera with minor brachiopod spat and calcispheres. Of particular stratigraphic interest (Section 2.5) are units within the Porty-yr-Aber Beds in which the large (up to 2 mm across) single chambered, sack-shaped foraminiferan Saccaminopsis is conspicuously abundant (Plate 70a). Fig.41c demonstrates the misleading high proportions of sparry calcite cement in these units almost wholly developed within the tests of these foraminifera.

The lithofacies locally contain a diverse coral/brachiopod macrofauna and displays a rich suite of trace fossils (Fig.42). Preservation of the various shelly fossils is variable. In general those within the shaly bands are severely crushed, whilst those observed within the limestones are well preserved, but have not always fully escaped the effects of compactional breakage.

Insoluble residues from the limestone of lithofacies I range between 20 and 30% by weight and consist mainly of clay minerals (illite, chlorite, kaolinite, illite/smectite mixed layer clays)

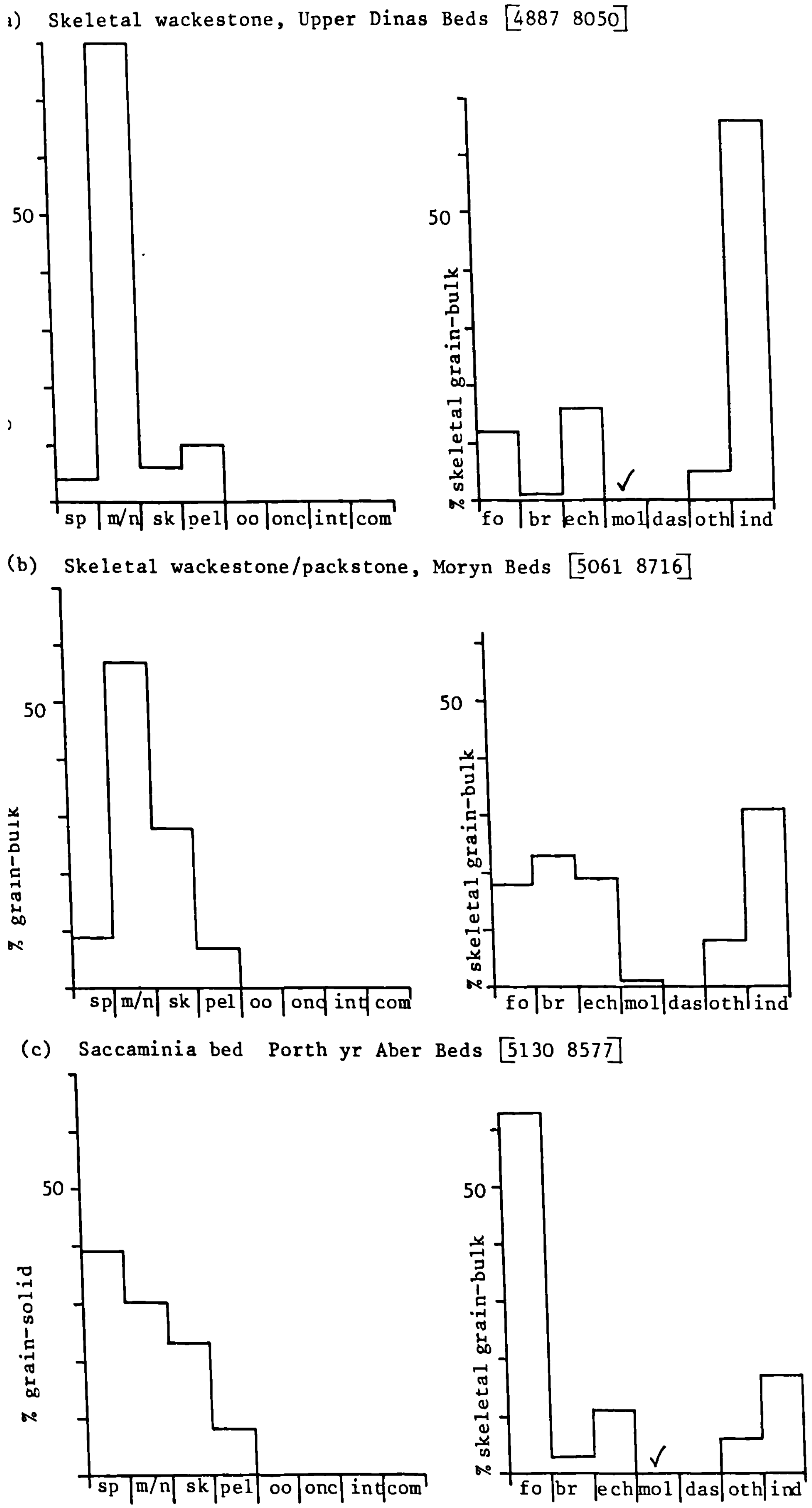


Fig.41 Compositional variations within Lithofacies I based on grain count analysis of selected samples (see Fig.44 for key).

Fig.42 Trace fossils drawn from cut block of skeletal wackestone/packstone bed, Lithofacies I, Porth yr Aber Beds [5130 8571].

with variable amounts of silica and pyrite. The effects of silicification within the lithofacies are evidenced in several ways. Scattered euhedral quartz prisms and beekitised fossils are ubiquitous, whilst nodular cherts are locally important in higher parts of the succession. Pyrite replacement of skeletal grains and particularly of the tests of foraminifera is widespread, the mineral also occurring as tiny scattered framboids.

(ii) Subfacies Ia : Shales with limestone nodules

Intercalated within the main lithofacies is a distinctive subfacies which consist dominantly of dark grey calcareous shale containing distinctive and often abundant limestone nodules which are clearly early diagenetic growths around burrow systems (Plate69). This subfacies is well developed in several of the Penmon cycles attaining a thickness of up to 3 m, e.g. T.B.4, and corresponds to the irregular-nodule beds of Nichols (1966). The nodules vary from simple cylindrical forms, up to 3 cms in diameter and 50 cms long, to branching and extensive sheet-like forms. The latter variety have grown around, and often preserve the detailed structure of the trace fossil *Zoophycus*. The shales which host the nodules contain much finely commuted skeletal debris as well as scattered Gigantoproductids, in life position, and crinoid stems up to 15 cms long. Thin sections of the nodules show them to be composed of argillaceous skeletal wackestone/packstone and therefore to be little more than calcite cemented parts of the host shales. Fine concentric lamination is often preserved and the skeletal debris is aligned parallel to the burrow sides.

(iii) Interpretation

The origins of limestone/shale alternations have long been the subject of debate, particularly the degree to which they reflect primary depositional layering as opposed to the effects of diagenetic segregation of $CaCO_3$. Hallam (1964) has reviewed the various mechanisms put forward in explanation of the closely comparable Liassic lithofacies. He concludes that most of the small scale alternations were of diagenetic origin, although these are thought to be superimposed upon broader primary layering as indicated by differences in faunal content, bioturbation effects and variations in limestone/shale ratio. Hallam argues that the diagenetically induced alternations were caused by 'rhythmic unmixing' of $CaCO_3$ and clay. He further contends that this took place soon after burial and certainly prior to compaction evidencing the well preserved nature of fossils within the limestones as opposed to those crushed within the shales. Wobber (1967) working on these same Jurassic strata interprets much of the nodularity of the limestone beds as due to sedimentary boudinage whilst low angle dislocations were, he suggests, caused by interstratal sliding; both these phenomena reflecting the effects of deep durial.

More recently Wanless (1979) has shown that many of the bedding characteristics of Lithofacies I are explicable with reference to pressure solution and are the typical results of compressive stress acting on argillaceous limestone. Dissolution of the $CaCO_3$ concentrates the non-carbonate impurities and leads to the formation of clay seams with nodular limestones the predicted end result. The implication of this work is that sequences of interbedded nodular limestones and shales may be generated during deep burial from an originally massive argillaceous limestone (see also Logan and Semeniuk, 1976).

In contrast to these diagenetic models has been the recognition of limestone turbidites and storm dominated carbonate shelf sequences with the realisation that carbonate/argillite rhythms may be hydraulically comparable with sequences of interbedded sandstone and shale (e.g.Meischner, 1964). Although the absence of grading and tractional sedimentary structures from Lithofacies I argues strongly against such interpretation, the effects of subsequent bioturbation (Fig.42) in obliterating primary structures should not be underestimated and such sudden event, particularly storm related, modes of deposition should perhaps not be discounted. Other mechanisms variously put forward in explanation of a primary alternation include oscillations in water depth, temperature, salinity, and rate of terrigenoclastic influx.

The predominance of micritic carbonate and terrigenous clay grade material indicates deposition in an essentially quiescent environment where such fines were allowed to accumulate without subsequent winnowing. Analogous modern day sediments have been recorded from the deeper (35-100 m), more distal parts of the Persian Gulf (Wagner and van der Togt, 1973) and from the Guatamala and Honduras shelf lagoon in water depths of 10 m and greater (Matthews, 1966; Ginsburg and James, 1974), whilst purely carbonate muds occupy extensive areas of the Bahama Banks (Bathurst, 1975). Terrigenoclastic fractions reflect deposition from suspension of material derived from older argillaceous strata exposed in the adjacent hinterland (Greenly, 1919 p.625) and of fluvial (c.f. Honduras shelf) and/or wind blown (Greenly, 1928; c.f. Persian Gulf) origin. Matthews (1966) has shown that carbonate muds in the Honduras shelf lagoon are also partly allochthonous, winnowed from high energy shelf edge environments, but also largely autochthonous, resulting from the disintegration of the tests of the indigenous shelly fauna. He further demonstrates that much of the sand sized shell debris within these muddy environments is also produced by the latter process.

In Recent settings such sediments are deposited below effective fair weather wave base although in this context the baffling effects of seaweeds may be important (Davies, 1970). Wave-base is clearly equivalent to depth of water, but the effects of shelf edge barrier systems in reducing the effective wind fetch over shelf lagoons may allow fines to accumulate in relatively shallow conditions. Thus on the Bahama Banks mud grade material is being actively deposited in waters under 4 m deep, whilst in sheltered coastal embankments of the Persian Gulf such sediments pass lateraly into indertidal flat deposits. Other evidence based on the thicknesses of the minor cycles (Section 5.6b) suggests that the lithofacies was unlikely to have been deposited in depths greater than 30 m and for the most part probably much less.

Parts of the lithofacies which yield a diverse fossil assemblage evidence a well oxygenated depositional environment populated by predominantly stenohaline taxa and therefore at, or near, normal salinities (Raup and Stanley, 1971).

(b) Lithofacies II : Skeletal packstone/grainstones

(i) Description

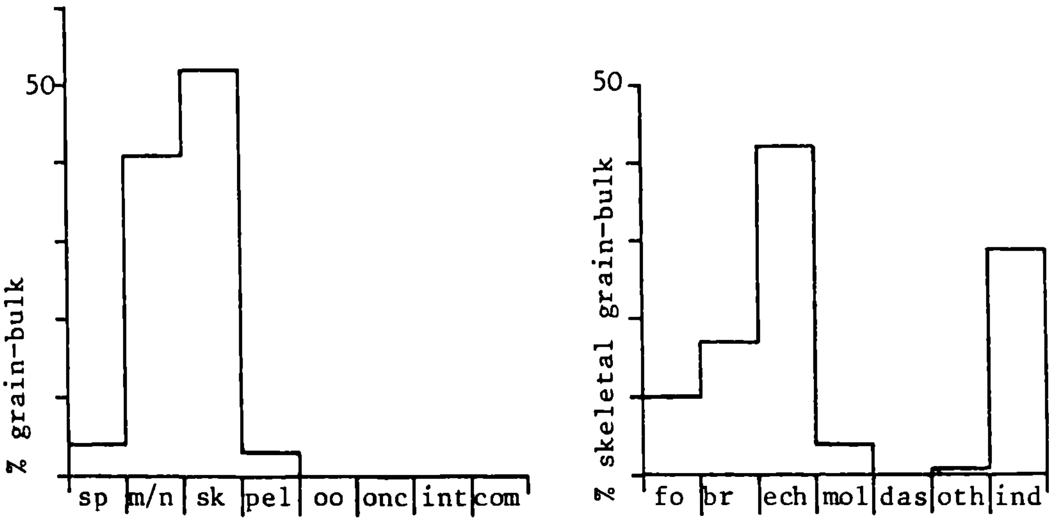
This lithofacies is volumetrically the most important within the Anglesey Dinantian, minor cycles in the Moelfre and Red Wharf Formations are constructed almost entirely from it, whilst the lithofacies characterises the upper portions of the cycles in the Traeth Bychan Formation. The lithofacies comprises thickly bedded dark tan to pale grey skeletal packstones, but interspersed within this background lithology are ill-defined, irregular clots, whisps and stringers of skeletal grainstone (Plate 71a) and the patchy distribution of the two textural types has demanded the use, in the field, of the umbrella term packstone/grainstone. Bedding within the lithofacies comprises widely separated (2 to 3 m) stylolitic shaly partings which occasionally thicken into rubbly zones, the so called 'stick beds' (Garwood, 1913 p.475) in which limestone casts of cylindrical and thalassinoidian type burrows are preserved within a shaly matrix. The limestone beds may be massive, but characteristically display pressure solution effects (Plate 70c) comparable with the stylobreccias of Logan and Semeniuk (1976). Cut blocks of these parts of the lithofacies display often pronounced colour mottling (Plate 71b). Many of the darker patches exhibit distinct linear and branching burrow forms or are developed in halos around macrofossils. No tractional sedimentary structures have been recorded from this standard lithology.

Distinct from the almost ubiquitous scattered pods of grainstone within the lithofacies are impersistent thin beds and lenses of skeletal grainstone and productid coquina which often exhibit low angle cross lamination and shell imbrication. Bases of such units are commonly sharp and lined by intraclasts and strongly suggest scour. Tops are gradational with the host packstone/grainstones and indicate mixing by bioturbation. Stringers of coarse shell and crinoidal debris have also invariably suffered the latter fate.

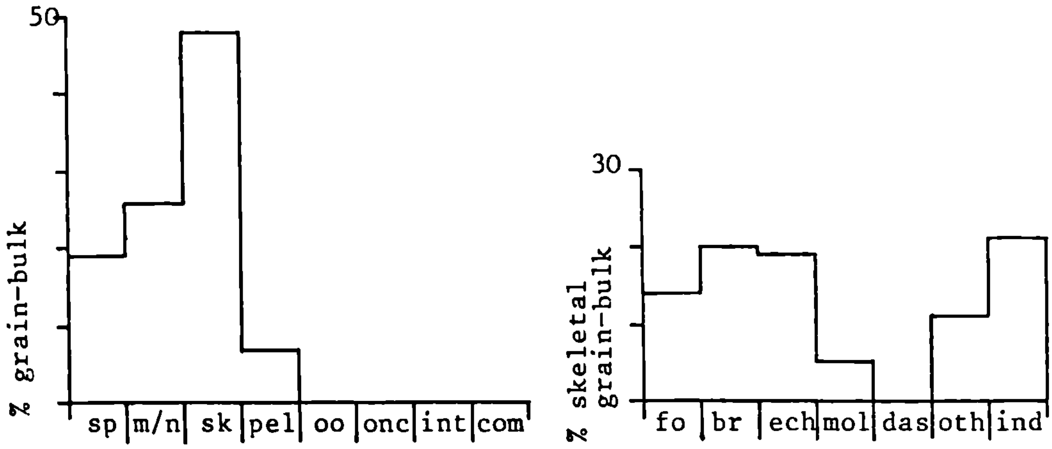
Units of the lithofacies rich in crinoidal debris also occur notably towards the top of the Eglwys Siglen Beds around Moelfre.

In thin sections of the background packstone lithology brachiopod and echinoderm debris and whole foraminifera dominate the recognised skeletal components (Fig. 43). Interdeterminate grains predominate over all, but most are clearly micritised skeletal

(a) Skeletal packstone, Porth yr Aber Beds [5130 8577]



(b) Skeletal packstone/grainstone, Upper Helaeth Beds [5112 8680]



(c) Grainstone lens in above

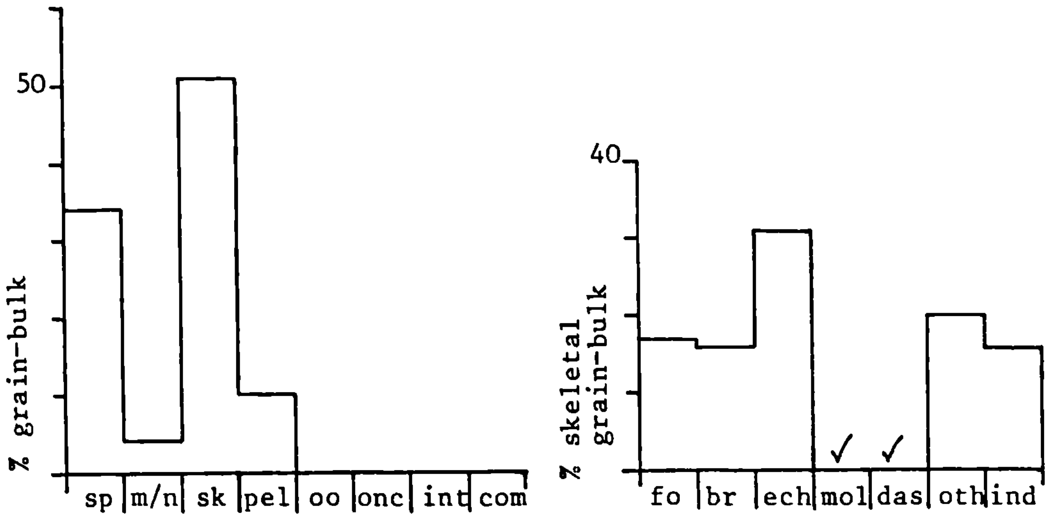


Fig.43 Compositional variations within Lithofacies II based on grain count analysis of selected samples (see Fig.44 for key).

debris. Minor mollusc, coral and bryozan debris occur whilst the distinctive fragments of trilobite carapace have also been recorded. The skeletal grains are typically both angular and poorly sorted ranging from micrite grade material to whole shells. Peripheral micritisation of skeletal grains is ubiquitous and in some of the thicker shell fragments discrete micrite filled algal borings are well seen.
The originally micritic matrix material has invariably undergone varying degrees of recrystallisation to neomorphic micro- and pseudospar.

Grainstone portions of the standard packstone/grainstone parts of the lithofacies contain the same foram/brach./echinoderm dominated grain assemblage as the packstones and appear therefore to be simply winnowed versions of the latter. Nor is the grain composition of the cross laminated grainstone and coquina lenses distinctive, although the standard skeletal grains are augmented by scattered intraclasts. Sorting in these latter units is improved and often bimodal with the grainstones forming a matrix to the whole shell coquinas.

Parts of the lithofacies intercalated with units of Lithofacies III often contain varying proportions of dasycladacean algal grains and/ or ooids. These grains are typically well rounded and well sorted and where they are particularly abundant the packstones of Lithofacies II exhibit textural inversion (Folk, 1962).

Macrofauna present within the lithofacies is variable. Sporadic solitary and colonial corals, gastropods and generally disarticulated productid brachiopods occur in the Moelfre Formation. Lengths of crinoid stem over 15 cm long recorded from the Lower Lookout and Eglwys Siglen Beds cannot have been transported far. Units of the lithofacies in the Brigantian Traeth Bychan Formation are more rewarding with abundant though scattered colonial corals and with both fasciculate and cerioid forms often attaining large diameters. Gigantoproductids augment the brachiopod fauna but again are often disarticulated. Insoluble residues are generally less than 10% by weight; units of the lithofacies gradational with Lithofacies I are likely to contain significantly higher proportions, the grainstone lenses much less.

(ii) Interpretation

The background packstone lithologies evidence the partial winnowing of fines such that grain support textures prevail, whilst the interspersed patches of grainstone demonstrate the periodic complete removal of fines. The present patchy distribution of the two textural types is thought to reflect mixing by bioturbation the mottling effects of which are clearly seen throughout the lithofacies. An initial intercalation of the two lithologies is therefore indicated, and the subordinate grainstone elements are envisaged, prior to bioturbation, as thin irregular spreads on a sea floor where for the most part skeletal packstones were accumulating.

The macrofauna of the lithofacies, whilst not always abundant, is dominantly of stenohaline forms and indicates an environment at or near normal salinities.

The lithofacies appears analogous therefore to the foram/mollusc sands and muddy sands widely developed on many modern carbonate shelves and platforms (Ginsburg and James, 1974) (Benthonic mollusca now occupy many of the ecological niches previously filled by Palaeozoic brachiopods, Raup and Stanley, 1971). The constituent skeletal grains of these sediments appear to have been largely derived from the in situ breakdown of indigenous shelly fauna. In the central parts of the Persian Gulf such muddy sands are gradational with axial argillaceous muds (c.f. Lithofacies I) and accumulate in water depths between 20 and 60 m (Wagner and van der Togt, 1973). Elsewhere within the Gulf, however, equivalent sediments are also being deposited in near shore settings within the more open parts of back barrier lagoons in water depths of less than 10 m (Purser and Evans, 1973). Subtidal muddy sands and packstone textures also prevail on sheltered leeward coasts e.g. east of Quatar Peninsular (op. cit.) and west of Andros Island (Schlager and Ginsburg, 1981).

These deposits are actively accumulating in environments at or around effective fair weather wave base or indeed well within the zone of wave activity; the latter where the baffling effects of sea grasses (e.g. Shark Bay; Davies, 1970) or the binding action of subtidal gelatinous algal mats (Bathurst, 1975) inhibit the movement of grains and prevent the complete winnowing of fines. On unprotected sea bottoms patches of grainstone are developed by more efficient winnowing in response to the temporary lowering of effective wave base during storm activity. On vegetated sea floors grainstones prevail in areas where sea grass cover is sparse or where subtidal mats have been destroyed by bioturbation or storm erosion. In these latter cases the sea floor often develops ripple-marks as tractional movement of the grains is allowed, but preservation of internal lamination is precluded by contemporary bioturbation (Hagan and Logan, 1974). In this context the erosively based, intraclast-lined and cross laminated grainstone and coquina lenses must represent deep scours within which sediments of sufficient thickness to escape the effects of burrowing benthos were deposited, possibly related to major storm events. Occasional vertical escape traces recorded from these units (e.g. Upper Helaeth Beds) are in accord with the latter interpretation. Units of the lithofacies containing dasycladacean algal grains and ooids and which often exhibit textural inversion evidence the proximity of higher energy shoal complexes from which such grains were winnowed (Folk, 1962; see Lithofacies III below).

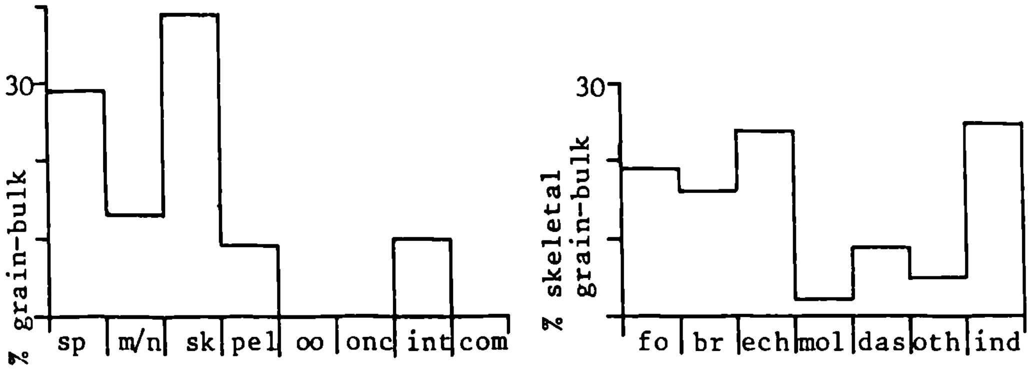
(iii) Subfacies IIa : Intraclast skeletal packstone/grainstones

This important subfacies is readily distinguished from the main parts of Lithofacies II. Where present it characteristically forms a basal phase to the minor cycles up to 2 m thick but is often impersistent pinching out against broad swells in underlying palaeokarstic surfaces e.g. at the base of the Pedolau Beds, Plate 72b. The subfacies contains abundant intraclasts of underlying, often calcretised, strata (Plate 73a,b) including skeletal fragments analogous to the lithoskels of Read (1974) (Plate 73c; see Section 5.3). Quartz pebbles and sand are locally present.

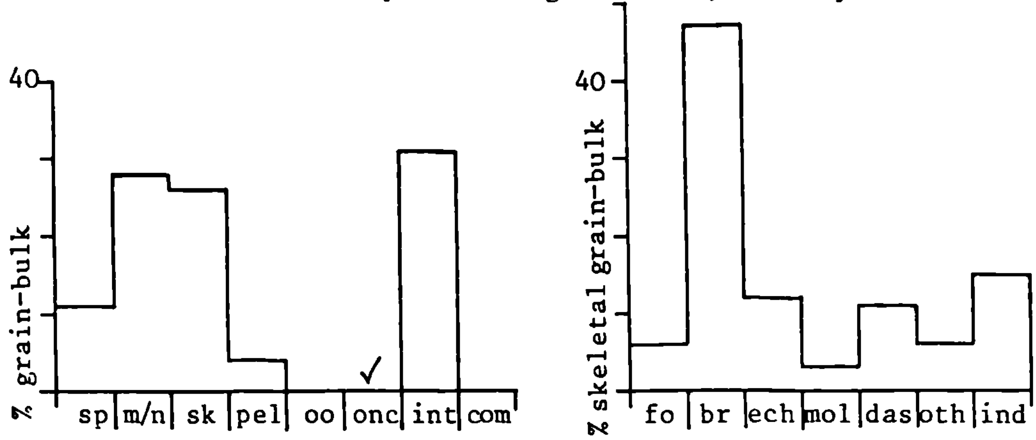
Units of subfacies IIa are generally coarser grained than those of the main Lithofacies and are locally coquinoid. Pore-filling calcite cements often predominate over micritic matrix material (Fig. 44) which may be only patchily developed; such textural mottling, often enhanced by colour, demonstrating the effects of bioturbation. The subfacies is texturally intergradational with grainstone units of Lithofacies III which also occupy a basal position within some minor cycles e.g. Flagstaff Formation. Matrix rich varieties of the subfacies IIa also occur.

The standard skeletal grain assemblage of Lithofacies II is augmented by often large fragments of dasycladacean algae (Plate 73b). Coquinoid units are rich in disarticulated productid valves bereft of spines and often imbricated. Rolled corals, typically robust

(a) Intraclast skeletal packstone/grainstone, Pedolau Beds [5069 8714]



(b) Intraclast skeletal packstone/grainstone, Porth y Rhos Beds [5140 8493]



KEY

sp	spar cement	fo	foraminifera
m/n	micrite/neomorphic spar	br	brachiopod
sk	skeletal grains	ech	echinoderm
pel	peloids	mol	mollusc
oo	ooids	das	dasycladacean algae
onc	oncoids	oth	others
int	intraclasts	ind	indeterminate
com	compound grains		

✓ = present, but less than 1%

Fig.44 Compositional variations within subfacies IIa based on grain count analysis of selected samples (including key to Figs.41, 43, 45 and 46).

cerioid forms or solitary types also occur, whilst oncolitic coatings around both shell fragments and intraclasts have been recorded.

(iv) Subfacies IIa : Comparison with modern analogues

The characteristic basal setting of units of subfacies IIa in many minor cycles leads to comparison with the transgressive lags which veneer extensive areas on modern carbonate shelves (Ginsburg and James, 1974). These are the residual deposits of high energy beach and nearshore environments which migrated landward during the Holocene rise in sea level. They are similarly rich in intraclasts of underlying strata whilst their shallow water ancestry is reflected in grainstone and coquinoid textures. Such deposits, however, are now out of equilibrium with their present deeper shelf setting and are actively being colonised by indigenous shelly and burrowing benthos (Logan et al, 1969). Bioturbation mixing of presently accumulating fines with the grainstone lags is likely to give the textural mottling observed in subfacies IIa.

A more detailed appraisal of such Recent transgressive units has been presented by Hagan and Logan (1974) based on observations in Shark Bay, Western Australia. They recognise a "Basal Sheet" which still floors deeper parts of coastal embayments, but which underlies prograding shallower water deposits. This 'Sheet' overlies Pleistocene limestones which suffered subaerial exposure and calcretisation during the last pre-Holocene low stand in sea level (Logan et al, 1970) and the upper surface of which is therefore closely comparable with the palaeokarstic surfaces on Anglesey. As with units of subfacies IIa Basal Sheet sediments locally up to 3 m thick pinch out against local swells in this underlying surface.

Upper parts of the Basal Sheet sequence comprise skeletal packstones and wackestones and coquinoid grainstone and reflect recent sediment accumulation following transgression. Such deposits may correspond to various other lithofacies recognised within the Anglesey Dinantian. Lower units of the Basal Sheet, however, compare more closely with subfacies IIa and comprise intraclast rich grainstone with an abundance of Read's lithoskels and also grains derived from Pleistocene soil profiles. These deposits vary from medium to coarse grained, contain a diverse assemblage of skeletal grains and are locally sandy. Hagan and Logan interpret these units as "formed by reworking of substrates at the strand" during transgression whilst subsequent extensive bioturbation by crustaceans and echinoids has led to homogenous fabrics and mixing with overlying matrix rich lithologies. Read (1974) records "mottled zones as thick as 1 m" at the contact between Basal Sheet sediments and overlying sea grass bank deposits (see Moelfre Formation, Section 5.6).

(c) Lithofacies III : Skeletal peloid grainstones

(i) Description

Skeletal peloid grainstones are particularly prevalent in the Careg-onen and Flagstaff Formations forming units over 8 m thick, but also commonly occur as a thin capping phase, generally less than 2 m thick, to minor cycles in the higher formations. The thicker units of the lithofacies in the lower parts of the sequence display well developed low angle and trough cross-bedding in cosets several metres thick (Plate 74c; Charts 1 and 2) separated by thin (< 50 cms) bioturbated packstone/grainstone horizons. Vertical and oblique escape traces are common within the cross bedded parts whilst within the often rubbly packstone/grainstone intercalations casts of cylindrical and branching thalassinoidian type burrows have been recorded. Trough cross bedded sets commonly exhibit internal grading, fining upwards from a coarse often rudstone textured toe-set phase in which oncoids, intraclasts and large fragments of the dasycladacean algae Koninkopora are readily observed (Plate 74b). The thinner units of the lithofacies in the Moelfre and Traeth Bychan Formations rarely exhibit large scale cross-stratification although low angle and planar lamination and current ripple cross-lamination are commonly observed (Plate 74a).

Coquinoid lenses of productid valves, rolled corals, Chaetetes, oncoids and intraclasts are a ubiquitous feature of the lithofacies occurring both in a basal lag position and scattered throughout. In the former occurrence such lithologies contain intraclasts of the underlying lithologies and evidence an erosive contact. Coquinoid lenses developed within the body of the lithofacies contain intraclasts of the host grainstones indicating early lithification and erosion; extraneous clasts of other lithologies are still present however.

The lithofacies erosively overlies most of the other lithofacies but gradational contacts particularly with lithofacies II are also common. Intercalation with the latter lithofacies also occurs but the grainstones of Lithofacies III commonly form the uppermost parts of the minor cycles and are overlain by palaeokarstic surfaces.

No in situ macrofauna has been recorded from the lithofacies.

Cut-block, peel and thin section studies of the lithofacies demonstrate the characteristically well rounded and well sorted nature of the constituent grains (Plate 75a,b). The standard skeletal grains, foraminifera, brachiopod and echinoderm, are all present though highly abraded, rounded and intensely micritised, the latter to the extent that the use of the term pelloid cannot be avoided (Fig. 45). Chambers and pores within skeletal grains are often plugged by micrite (Plate 75b). Significant additions to the allochem assemblage are grains of dasycladacean algae, intraclasts and grapestone-like compound grains. The former algal grains are locally so abundant as to occlude the other skeletal elements and form almost pure dasycladacean grainstones (Plates 74c,75b). These units appear to be of very limited extent however, occurring as scattered pods or lenses within the more normal skeletal pelloid grainstones. Other parts of the lithofacies are locally rich in crinoidal debris and constitute encrinite grainstones (Fig.45c; Wilson, 1975 p.65). Superficial oolitic coats and scattered well formed ooids demonstrate the gradational relationship with the oolitic lithofacies IV described below.

The grains are cemented by sparry calcite. Thin pallisade fringe cements are commonly developed on many grains, whilst remaining pore space is filled by coarse blocky spar. Insoluble residues from the lithofacies are less than 1% by weight.

(ii) Interpretation

The grainstone and locally rudstone textures, rounding and sorting characteristics and tractional sedimentary structures record deposition under sustained high energy conditions. Intercalated bioturbated packstone horizons, however, suggest possibly prolonged periods of quiescence during which micrite grade material was allowed to accumulate and burrowing benthos colonised the substrate.

Recent dasycladacean algae live in water depths of up to 30 m but thrive in shoal areas under 5 m deep (Ginsburg et al, 1971). Units predominantly composed of these algal grains are thought to

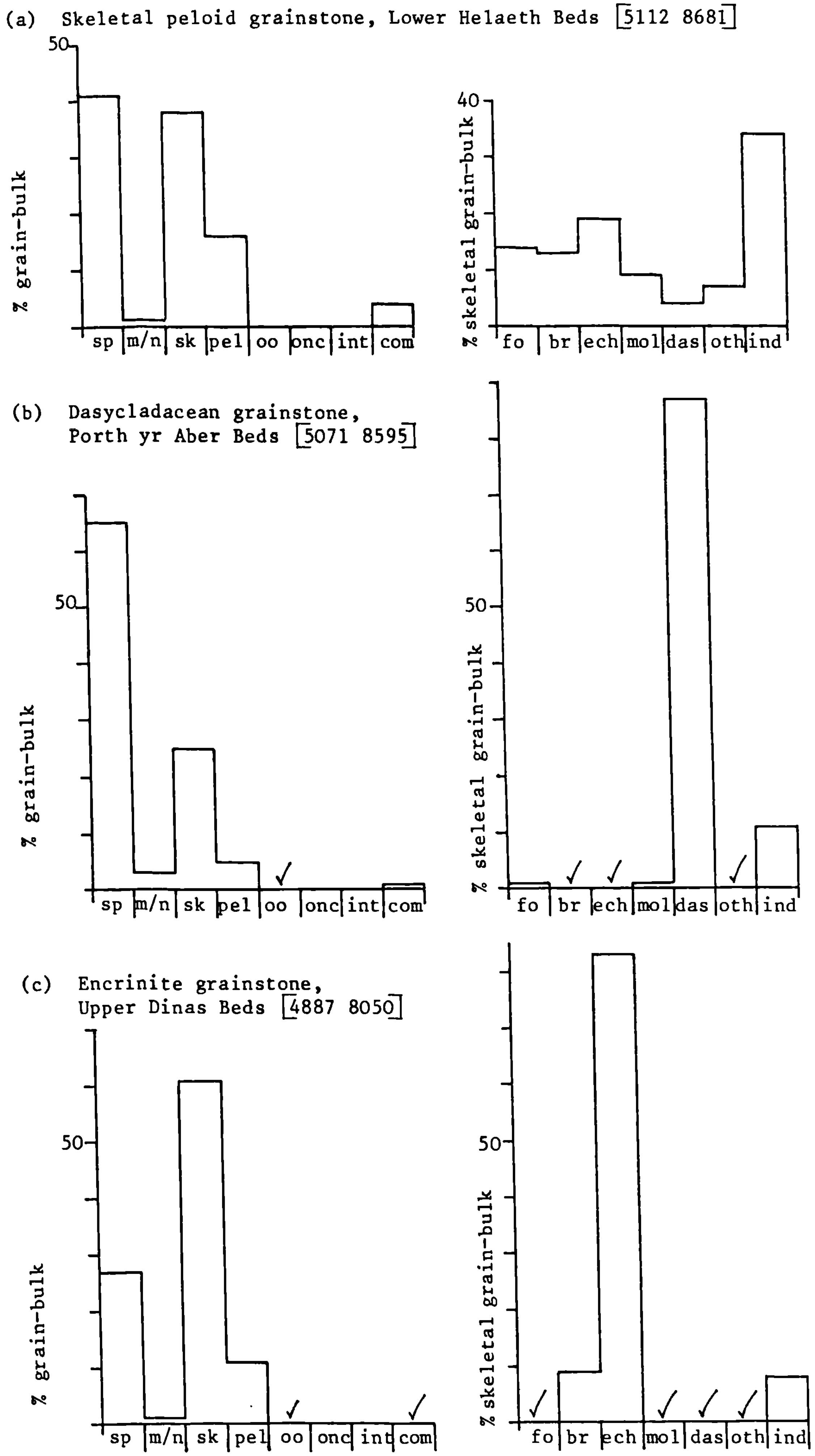


Fig.45 Compositional variations within Lithofacies III based on grain count analysis of selected samples (see Fig.44 for key).

reflect deposition within the latter shallow water setting (Wilson, 1975) whilst the paucity of bioclasts derived from other shelly taxa (Fig.45b) may indicate locally restricted conditions (Flugel, 1982). Conversely the standard parts of the lithofacies containing a diverse assemblage of skeletal grains (Fig.45a) suggest derivation from areas of skeletal production at or about normal salinities.

Encrinite grainstones require less energetic conditions for their accumulation, a reflection of the porous nature and relatively low specific gravity of even quite large echinoderm plates (Bathurst, 1975 p.50). Here the predominance of one type of skeletal grain, derived from a stenohaline group, is likely to reflect proximity to the sites of skeletal production probably areas of the sea bed colonised by dense crinoid meadows.

The grapestone-like aggregate recorded from the lithofacies demonstrate the early submarine cementation of adjacent grains during periods of immobility. On the Great Bahama Bank the binding of the grains within a subtidal algal mat most facilitates this process (Bathurst, 1975 p.316). Grapestones are thought to characterise areas of reduced skeletal production and to form in "very warm, shallow water with only moderate circulation" (Wilson, 1975) conditions which favour the ready precipitation of submarine cements. Unlike the Bahamas, however, the formation of grapestone aggregates within the Anglesey succession was of only local importance and they comprise less than 10% of the constituent grains (Fig.45)

The grainstones of Lithofacies III are closely comparable with modern carbonate sands so characteristic of shelf edge shoal environments, but which are also important along landward shore lines forming barrier bars and beaches (Ball, 1967; Purser, 1973;

Wilson, 1975). These Recent sediments are forming mainly in shallow water sèttings within the zone of constant wave activity and where tidal currents may also be important. Subaqueous shoals, however, commonly build up above sea level and may be capped by beaches and storm ridges, whilst on landward margins these may in turn give way to extensive dune fields of wind blown carbonate sand. These latter settings serve to remind us that cross bedded grainstones may also accumulate under largely subaerial conditions with perhaps a strong aeolian component (Ball, 1967; Logan et al, 1970).

Carbonate shoals embrace a range of subenvironments including tidal and storm surge channels, tidal deltas and spillover lobes, whilst barrier and landward beaches may exhibit features comparable with sandy shorelines including nearshore bars, rip-channels and ridge-runnel systems. A variety of settings therefore in which the tractional sedimentary structures observed in Lithofacies III may have been developed. More detailed observation is required however before meaningful comparison can be attempted. Trains of megaripples which commonly traverse carbonate shoals are often only mobile during storm activity and for most of the time remain stationary with minor reworking of the crests by daily wave and tidal processes (Bathurst, 1975 p.121). The fines which are thus allowed to accumulate within the intervening troughs may correspond to the bioturbated packstone units recorded from Lithofacies III.

(iii) Subfacies IIIa : Thinly bedded skeletal grainstones with shaly partings

This subfacies forms an important component in many of the minor cycles in the Flagstaff Formation forming units up to 2 m thick. It may be both underlain and overlain by units of the main

Lithofacies, but may also be overlain by units of Lithofacies II. Further minor occurrences have been recorded from the Traeth Bychan Formation. The Lithofacies consists of thin (up to 10 cms) parallel sided to lenticular beds of skeletal grainstone and coquina (Plate 76a). The individual beds may be separated by thin shale laminae or thicker (up to 25 cms) shale rich rubbly zones in which casts of cylindrical burrows are often evident. The Lithofacies is unusual in its apparent susceptibility to recrystallisation. Enterolithic clots of microspar abound within the shaly intercalations and invade, and locally totally replace the limestone beds e.g. at Pedolau (Plate 75c). These effects, which impart the often conspicuous rubbly appearance to the subfacies, are thought to result from vadose diagenesis during periods of emergence and palaeokarst formation and are discussed more fully in Chapter 3, Section 4_g.

In their less altered state the limestone beds commonly display cross lamination (Plate 76b) often disrupted by bioturbation. A vague fining upwards within some beds has been noted whilst cut and fill structures with one bed eroding down through underlying units have also been observed. Whole shell coquinas of chonetid brachiopods are common in the Flagstaff Formation whilst equivalent units in the Brigantian are often rich in rhynchonellids. The shells within these coquinoid units are often still articulated and the latter more globose rhynchonellids occasionally display rotated geopetal sediment fills.

Cut-block and thin section examination of the skeletal grainstone beds reveals the same assemblage as the main part of the lithofacies although again one grain type may predominate over the others and beds rich in crinoidal debris (encrinites) or productid shell hash or even fragments of fasciculate coral have been observed. The grains are generally moderately well sorted, often bimodal in the coquinas; rounding varies from good to poor. In parts of the beds which have not suffered recrystallisation the grains are set in sparry calcite cements as observed in Lithofacies III.

(iv) Interpretation

The grain composition of the subfacies, and its often intimate field relationships, firmly links it with the thicker, cross-bedded units of the main Lithofacies. In some respects the thinly bedded nature of the subfacies is reminiscent of the stylobedding of Logan and Semeniuk (1976; compare their fig.30 with Plate 75c). Yet the lateral continuity of thin units of the subfacies e.g. in the Flagstaff Formation (Fig. 52 and the occurrence of burrow forms within the shales argue strongly against an origin simply through pressure solution. The intercalated shaly laminae and locally thicker shale rich rubbly zones indicate an environment in which terrigenous fines were allowed to settle out of suspension (c.f. Lithofacies I). The thin coquina and cross laminated grainstone beds therefore record periodic high energy influxes which were subject to subsequent bioturbation. These features are closely comparable with those displayed by lower shoreface deposits of siliciclastic regimes where coarser sand material is introduced during storm activity and then reworked by indigenous benthos (Reineck and Singh,1975). On this basis the units of subfacies IIIa are thought to record deposition on the lower flanks of the main carbonate shoals as represented by Lithofacies III. The grainstone beds thus represent material swept off the shoal crests during storms and redeposited on the deeper flanks by storm surge ebb currents.

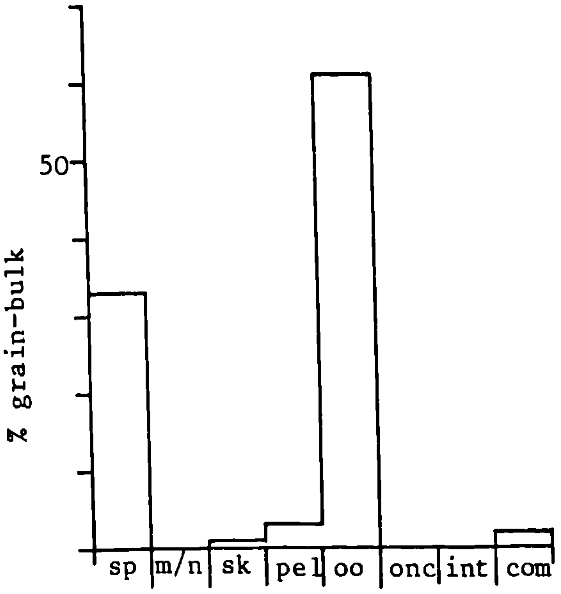
(d) Lithofacies IV : Ooid, skeletal grainstones

(i) Description

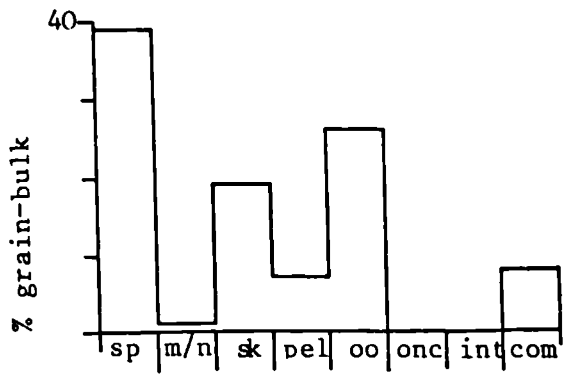
Oolitic limestones have previously been thought of as uncommon in the Anglesey succession (Greenly, 1919 p.606) only two examples having been recorded; the Edwen oolite from the Straitside section (op. cit. p.649) and on the island of Ynys Dulas (Mitchell, 1964 p.64). In fact thin laterally discontinuous oolitic units are quite common towards the tops of minor cycles where they occur intimately associated with the skeletal peloid grainstones of Lithofacies III. The environmental significance often attached to oolitic deposits has prompted their separate description and interpretation.

Oolitic units are often cross-laminated whilst thicker units of the Lithofacies in the Craeg-onen Formation exhibit large scale cross bedding similar to that observed in Lithofacies III. In thin sections of the Lithofacies ooids (see below) may comprise up to 90% of the constituent grains but generally much less, diluted by varying amounts of associated skeletal grains, peloids, grapestones and intraclasts (Fig.46;Plate 77c). The skeletal grain assemblage compares closely with that of Lithofacies III with dasycladacean algae well represented. Oolitic limestones are characteristically well sorted, units with a high proportion of ooids particularly so. The ooids are by definition well rounded, and rounding of the associated grains is, in general, also good although angular skeletal particles do occur (Plate 77) however and textural inversion of Folk (1962, Fig.7) is therefore exhibited. The latter particles are often of dasycladacean algae and may reflect the ease of breakage of such grains across their large pore systems. The calcite cements of this grainstone lithofacies exhibit the same fabrics and

(a) Ooid grainstone, intraclast within palaeosol, top Upper Helaeth Beds [5150 8685]



(b) Ooid skeletal grainstone, Porth y Rhos Beds [5149 8572]



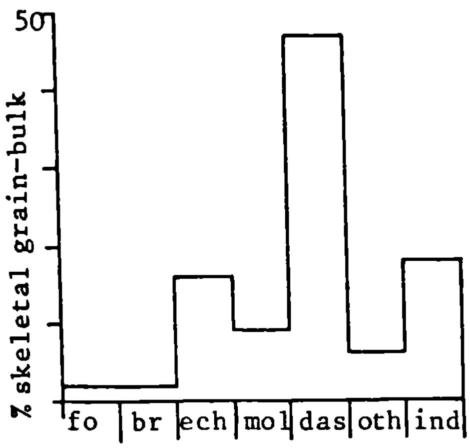


Fig.46 Composition variations within Lithofacies IV based on grain count analysis of selected samples (see Fig.44 for key)

composition variations as Lithofacies III.

The nature of ooid grains was not dealt with previously in Section 6.3, since the ensuing discussion on oolitic deposits demands their particularly detailed description and which may now be given. The ooids are small with an average diameter of 0.3 mm. Maximum diameters rarely exceed 0.5 mm (and then only where larger skeletal grains exhibit thin oolitic coats). Each ooid is composed of a nucleus with a surrounding cortex of radial fibrous calcite with concentric lamellae of dark micrite (Plate 77b). The nuclei are of the various types of grains which occur uncoated within the lithofacies (Fig.46). They vary in size from 'pin-point' nuclei a few microns across to the larger grains mentioned above. Ooid cortices correspondingly vary in their development from thick and multilayered to thin single coats only 10μ thick (superficial ooids) average thicknesses being about 100μ. The cortices are usually divided into three or four layers of prominent micrite lamellae, around 5μ thick, but under high magnification (x50) a whole series of increasingly finer lamellae can be observed with probably many more beyond the resolution of the microscope.

Within the ooid cortex the relationship between the radial fibrous calcite and the concentric micrite lamellae varies with the thickness of the lamellae. Most fine lamellae are in fact simply zones of inclusions in continuous calcite fibres, but where thicker and more substantial fibres terminate against them. The calcite fibres also tend to group together into larger clusters termed rays (c.f. Kahl, 1974 and Halley, 1977) and these commonly pass through even some of the thickest micrite lamellae, which are again reduced to a diffuse zone of inclusions. These fabrics appear almost identical to those described by Halley from recent aragonitic ooids (Halley, 1977 fig.6a) and by McGannon (1975) for calcitic fluvial ooids from the Pleistocene of Texas.

Multiple ooids where usually two ooids have joined and have the same outer oolitic coats (Plate 77b) are not uncommon. Asymmetric ooids, with the radial fibrous calcite preferentially developed on only one side of the nucleus, occur rarely. No broken ooids have been recorded.

Also included here as ooids, although some workers (Friedman et al, 1973 and Rubin and Friedman, 1977) advocate the use of the term spherulites, are grains made up of radial fibrous calcite the same size as normal ooids, but with indistinct micritic nuclei and with poorly developed concentric lamellae.

Ooids exhibit varying degrees of micritisation, those from units rich in micritised bioclastic grains are similarly extensively micritised.

(ii) Ooid Diagenesis : a review

Most modern ooids are composed of concentric lamellae of tangentially arranged aragonite needles (Bathurst, 1975). Ancient ooids are invariably of calcite with a radial fibrous structure (Sandberg, 1975). The generally accepted theory, applying uniformitarianism, is that ancient ooids were originally concentric and of aragonite and acquired their radial fibrous calcite form during diagenesis. Recent studies of ooid fabrics (Kahl, 1974; Sandberg, 1975 and Halley, 1977) pointing out the occurrence of primary radial fibrous aragonite ooids have enabled more refined diagenetic and environmental interpretation, but paradoxically have also raised serious question marks against the validity of the modern analog.

The mechanism proposed by Shearman et al (1970) for the transformation, in ooids, of aragonite to radial fibrous calcite has gained widespread acceptance. They suggest that piece-meal dissolution of originally tangentially arranged aragonite needles leaves a partly supported framework of concentric shells of the organic matter present within ooids. These then act as templates for the growth of radial crystal of calcite.

The major problem with this process and similar ones put forward by Carozzi (1962) and Loreau (1969) is in explaining the difference in diagenetic behaviour of ooids to that of skeletal aragonite. With skeletal aragonite grains inversion to calcite, even where neomorphism has taken place (concomitant dissolution and precipitation) and organic layers are preserved, always gives an equigranular sparry mossaic. Shearman et al (1970) clearly felt that the organic lamellae in ooids was of great significance here. Mitterer (1972), however, has shown that the protein composition of organic matter in modern ooids is the same as that in skeletal carbonates and they should, therefore, be expected to react in a similar way during diagenesis!

Investigations by Kahl (1974) on ooids from the Great Salt Lake, Utah, emphasized the occurrence of modern ooids with a radial fibrous structure and their association with hypersalinity. Such ooids are not uncommon, they have been recorded from Laguna Madre, Texas (Rusnak, 1960 and Freeman, 1962), the Gulf of Aquba (Friedman et al, 1973) and the Persian Gulf (Loreau and Purser, 1973) all hypersaline environments. Kahl (1974) showed the radial fibrous fabric to be primary and felt that similarity of textures in ancient ooids showed their radial fibrous fabric to be original also. He accordingly questioned the comparison of modern marine ooids with their dominantly concentric structure with ancient radial fibrous ones, modern equivalents of which appear to be restricted to hypersaline environments. In his discussion on diagenesis he suggests some syndepositional recrystallisation of the radial fibrous fabric took place to produce the bundles of larger aragonite crystals, called rays. He concludes that these primary and syndepositional aragonite radial fibrous fabrics underwent some forms of paramorphic replacement (Friedman, 1964 and Sanders and Friedman, 1967; inversion of one polymorph to another with retention of the original fabric) to give the radial fibrous calcite composition of ancient ooids.

Halley (1977) followed Kahl in his interpretation of ooid diagenesis and the correlation of radial fibrous fabrics with hypersalinity. He, however, seems to imply that modern marine ooids with tangentially arranged aragonite needles may still invert to radial fibrous calcite along the lines proposed by Shearman et al (1970) and that infact the only good evidence for primary and/or syndepositional radial fibrous fabrics (and, therefore, he felt hypersalinity) is syndepositional breakage along such a fabric. In support of this he shows that in both the Recent and ancient broken ooids and evaporitic deposits occur together.

Shearman et al (1970) also recognised early breakage along radial fibrous structure but in their examples this was clearly caused during compaction after the diagenetic transformation which they envisage had been completed.

It is strange that Halley does not take account of the fascinating paper by Sandberg (1975) who argues that not only is the radial fibrous structure of most ancient calcite ooids primary, but so also is their mineralogy! Paramorphic inversion of aragonite to calcite is shown to be as equally implausible a method of preserving radial fibrous fabrics (Sandberg, 1975 p.519) as that put forward by Shearman et al is for creating them, and for the same reason. Why should aragonitic ooids whether radical fibrous or concentric behave radically differently from aragonite skeletal grains? Without a satisfactory diagenetic mechanism the original calcite mineralogy of ancient ooids must be accepted. The strongest evidence in support of this idea comes from ooids in Pleistocene limestones from Florida. These ooids, which it is probably safe to assume, were originally identical to their modern relatives close by and therefore aragonitic, are now preserved in equigranular sparry calcite often neomorphic with ghosts of original concentric lamellae still present (Sandberg, 1975 fig.17). In short, they have acted in an identical way to contemporary aragonitic shell fragments. Primary radial-fibrous aragonite fabrics would not survive diagenesis and ancient ones must, therefore, have been originally calcitic.

(iii) Implications for the chemistry of ancient seas

Whey then are present day marine ooids of aragonite and ancient ones calcitic?

It is well known that the high Mg^{2+} content of modern seas ($Mg^{2+}/Ca^{2+} = 5/1$) prevents the growth of non-skeletal low-Mg calcite in favour of aragonite and high-Mg calcite (Bathurst, 1975). Sandberg (1975) argues quite convincingly that the most likely way to explain the formation of non-skeletal low-Mg calcite in ancient seas i.e. in ooids is to have had higher Ca^{2+} concentrations ($Mg^{2+}/Ca^{2+} < 2/1$ would have favoured the direct growth of calcite from sea water).

Sandberg suggests a reduction in the Ca^{2+} content of the seas during the Mesozoic due to massive extraction by calcareous planktonic organisms (foraminifera and coccoliths) and that the Mg^{2+}/Ca^{2+} ratio reached the critical levels to inhibit calcite growth in the early Caenozoic. He further suggests that as a consequence of lower Mg^{2+}/Ca^{2+} ratios in the past many calcareous algae would precipitate calcite skeletons, not aragonite as they do today, and that the muds to which these contributed on disintegration would also, therefore, have been calcitic in character.

(iv) Environmental significance of ooids and oolitic deposits

If we accept the primary nature of radial fibrous calcite in ancient ooids (Sandberg, 1975) then do many of our recently acquired ideas (Rusnak, 1960; Kahl, 1974; Loreau and Perser, 1973; Halley, 1977) relating various aragonitic fabrics (tangential to radial fibrous) in modern ooids to environmental factors become immediately obsolete? Sandberg (op. cit.) seems satisfied that the gross environmental conditions of ooid formation have always remained the same, i.e. shallow agitated marine waters, supersaturated with $CaCO_3$ (see Bathurst, 1967 p.448). Certainly, the common occurrence of cross bedding, often bipolar and the palaeogeographic setting as shoals etc. of ancient oolitic deposits (Heckel, 1972) would seem to support this contention. And yet, if this is the case, ancient radial fibrous fabrics formed in the same environments where tangentially arranged aragonite needles grow today. Sandberg implies that this change of fabric is simply a function of the change in mineralogy related to the different Mg/Ca ratios of ancient seas. Yet how does this account for the apparent differences in physical properties of the two fabrics?

Why, for instance, do modern radial fibrous aragonitic ooids from hypersaline environments exhibit such a high degree of breakage

(Halley, 1977) whilst ancient fully marine forms, of calcite, do not? The radial fibrous structure certainly appears to be an inherently weak one. Of great significance here, too, is the record of syndepositional breakage among primary calcitic fluvial ooids described by McGannon (1975) from the Pleistocene of Texas. This shows the difference cannot be explained simply in terms of mineralogy.

Recent theories suggest that the tangential arrangement of aragonite needles in modern ooids is due to the mechanical modification of embryonic needles growing with an original random or, in fact, radial fibrous distribution (Rusnack, 1960 and Loreau and Purser, 1973). Were calcite fibres less easily reorientated in this way?

In short, can the different fabrics of recent and ancient marine ooids be simply the result of differences in mineralogy, or must we revert to some apriori reasoning to account for the dominance of radial fibrous fabrics in the past? Are, perhaps, modern radial fibrous aragonitic ooids more indicative, not of salinities, but of the mechanical factors influencing the growth of this fabric than we are prepared to accept?

Kahl (1974) suggests that "agitation may be as important or more important than hypersalinity . . ." in controlling the type of fabric the ooid cortex takes. He felt that radial fibrous fabrics were indicative of more poorly agitated environments. Halley (1977) states that radial fibrous aragonitic ooids from modern hypersaline environments form "without the aid of severe agitation". Loreau and Purser (1973) as part of their theory of mechanical modification of ooid fabrics discussed above, show that ooids from the Persian Gulf with radial fibrous fabrics only occur in sheltered, quiet water lagoons. Freeman (1962) describes similar "quiet water ooids from Laguna Madre, Texas". Ancient oolitic sediments, however, including some from the British Lower Carboniferous e.g. Gully Oolite (Murray and Wright, 1971) with abundant cross bedding and cross lamination seem demonstrably high energy deposits, although the question of sites of formation as opposed to sites of accumulation may be significant here.

Further discussion on the various chemical and bio-chemical processes which may or may not control ooid growth and fabric (see Bathurst, 1975) is beyond the scope of this thesis, but in any event it is clear that the ancient oolitic deposits should not be subject to crude comparison with their modern counterparts. Indeed, until the processes which operated in what seem likely to have been chemically different ancient seas (Sandberg, 1975) are better understood environmental analysis not only of ancient oolitic, but ancient carbonate deposits in general must be regarded as incomplete.

(e) Lithofacies V : Calcite mudstones

(i) Description

At outcrop and in hand specimen the lithofacies is observed as characteristically white weathering porcellaneous calcite mudstones often displaying well developed 'birdseye' structures (Plate78a; Shinn , 1968; Deelman, 1972). On fresh surfaces colours range from dark grey to fawn, the darker varieties giving a foetid smell when newly broken. The lithofacies constitutes an important element of the Careg-onen Limestone Formation (Section 3.4b) occurring in beds up to 1.5 m thick separated by carbonaceous shale bands and intercalated with cross bedded, skeletal and

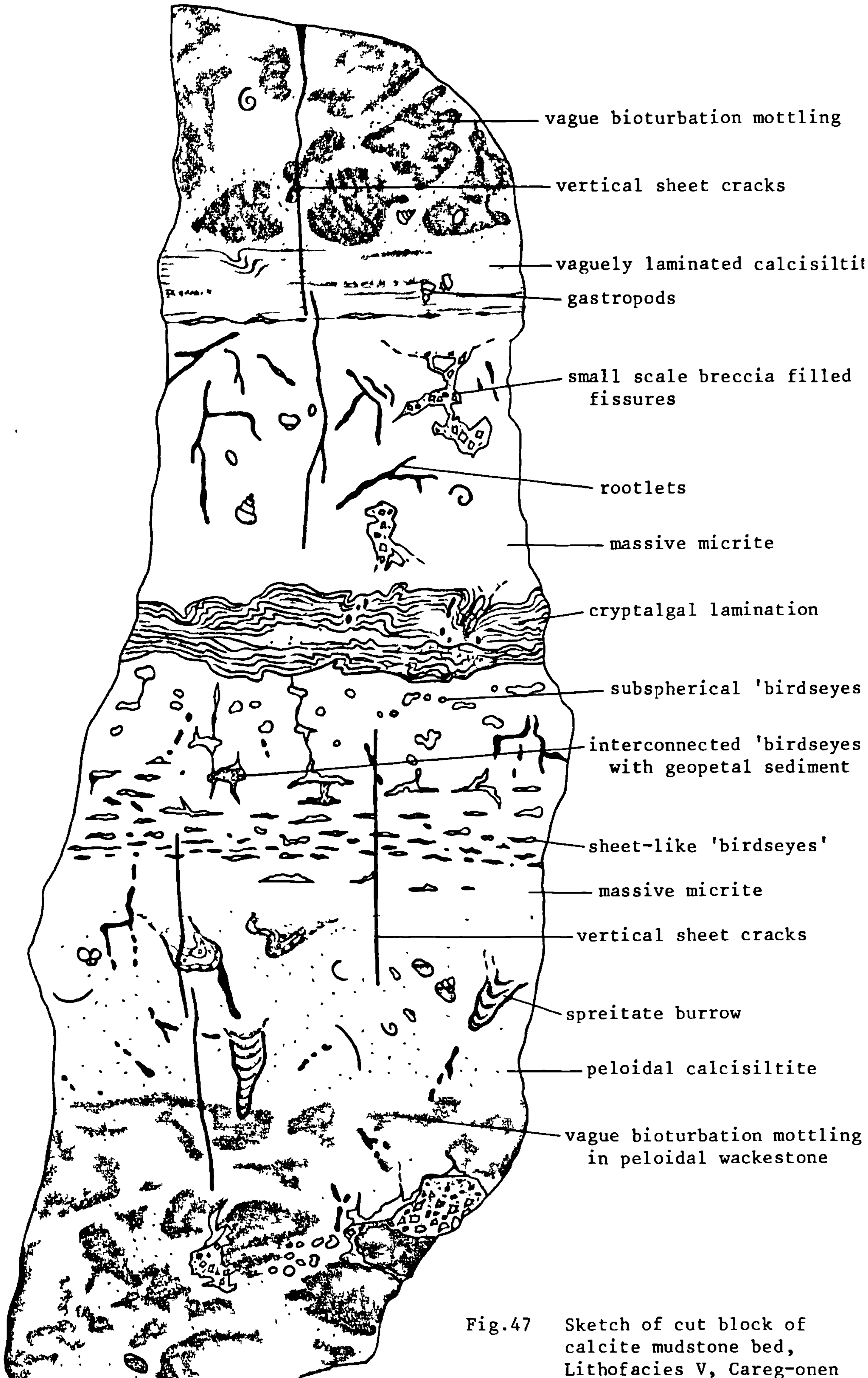


Fig.47 Sketch of cut block of calcite mudstone bed, Lithofacies V, Careg-onen Formation, Tandinas Quarry.

oncolitic grainstones (Plate 79b). Occasional beds exhibit well defined, planar or crinkly cryptalgal lamination. No domal or columnar stromatolites have been recorded. Desiccation cracks (Plate 78b) and carbonaceous rootlets have been observed, whilst turbinate gastropods are ubiquitous and comprise the only macrofauna of note (Fig.47).

Cut block, peel and thin section studies show the lithofacies to be predominantly composed of micrite patchily recrystallised to microspar, but with scattered ostracods and calcispheres (Plate 79c). The ostracods are smooth valved and occur both articulated and disarticulated, the former often enclosing geopetally arranged internal sediment overlain by sparry calcite. Where these micro-fossil components are particularly abundant the rock type approaches a wackestone type texture. Vague micritic peloids, often concentrated in bands and associated with bioturbation mottling, are also commonly observed.

The 'birdseye' structures or 'fenestrae' (Tebbutt et al, 1965) which characterise the calcite mudstone lithofacies occur on all scales and exhibit a variety of forms. Irregular sheet-like forms (Plate 78c) up to 3 cms long often impart a strong horizontal fabric to the rock. Sub-spherical varieties (Plate 79a) range from pin-prick size (readily confused with calcispheres) to 2 or 3 mm in diameter, whilst larger, irregular stromotactoid vugs up to 4 or 5 cms scross also occur. The fills are generally of sparry calcite, but flat topped internal sediment lining the bases of the structures is not uncommon. Vertical and oblique spar filled cracks and gashes abound and often form interconnections between the fenestrae (c.f. Fischer, 1964). Birdseye structures may be scattered throughout a calcite mudstone bed or be concentrated in a particular part generally towards the top (Plate 78a). The various types of birdseye may occur together with sheet-like fenestraemerging into sub-spherical structures or in other cases different parts of the bed may be characterised by different fenestrae types.

Other features of the lithofacies include rare lath shaped structures of sparry calcite, less than 0.5 mm long. These occasionally criss-cross one another in style typical of the twinning in gypsum crystals and are thought to represent pseudomorphs of the latter.

(ii) Interpretations

The micritic nature of the lithofacies indicates a depositional environment characterised by low energy conditions, whilst the occurrence of interbedded grainstone lithologies demonstrates that higher energy conditions periodically prevailed. The presence of rootlets and desiccation cracks suggests an environment subject to periodic emergence at or around depositional base level. The lithofacies is thought to record deposition on peritidal carbonate mud flats, the features it displays comparing closely with analogous modern sediments described by Shinn et al (1969). Such an interpretation is consistent with the restricted ostracod/calcisphere microfauna (Walker and Laport, 1970) and with the occurrence of cryptalgal lamination and gypsum pseudomorphs (Logan et al, 1970). Modern intertidal carbonate flats are characterised by abundant browsing cerithid gastropods which bear a strong resemblance to the small turbinate forms preserved within the Anglesey strata.

Birdseye structures are thought to be strongly indicative of the supratidal environment (Shinn, 1968) although they have been recorded from other settings e.g. lacustrine muds (Reineck and

Singh, 1965) and laminar calcretes (Section 3.4f). Those observed within the Anglesey calcite mudstones, however, are closely comparable with the ones described by Shinn from Recent supratidal carbonate flats and which he was able to reproduce experimentally. The flattened sheet-like forms (planar vugs) result from the repeated wetting and drying of the sediment and the resultant internal slippage and buckling caused by alternating expansion and shrinkage. Sub-spherical fenestrae are formed by gas bubbles evolved from decaying organic matter within the sediment, their retention and preservation being favoured by the partially consolidated nature of supratidal muds.

The occurrence of birdseye structures particularly towards the tops of calcite mudstone beds suggests that perhaps only these upper parts were laid down within the supratidal zone. Bioturbated and pelleted parts of the lithofacies may reflect deposition under more frequently wetted intertidal conditions or within surface ponds (Shinn et al, 1969).

5.6 FACIES ANALYSIS

Within the context of the North Wales Dinantian shelf lagoon the Anglesey sequence is likely to reflect deposition in a complex mosaic of depositional environments. The deposits of these various environments have been resolved into the five principal lithofacies mainly on the basis of textural parameters which are not environmentally unique. The more precise interpretation of depositional environment rests with an appreciation of the sequences and associations of lithofacies within the various formations. Analysis of the various lithofacies relationships and associations which comprise the formations has been achieved using semi-quantitative techniques outlined by Harms et al (1975). This has allowed the construction of facies diagrams (e.g.Fig.48) illustrating the observed patterns of lithofacies within the various formations and showing the main sequential trends. Thus, whilst there is considerable overlap between the formations in the lithofacies they contain, each formation is in general characterised by a particularly prevalent lithofacies association. Indeed it is this property which makes the formations lithostratigraphically distinctive. Facies analysis of the higher formations is made more complicated by their minor cyclic character, the theoretical constraints of which are discussed in Section 5.6(b). The Careg-onen Formation is not subdivided into minor cycles and is discussed separately and first.

(a) Careg-onen Formation

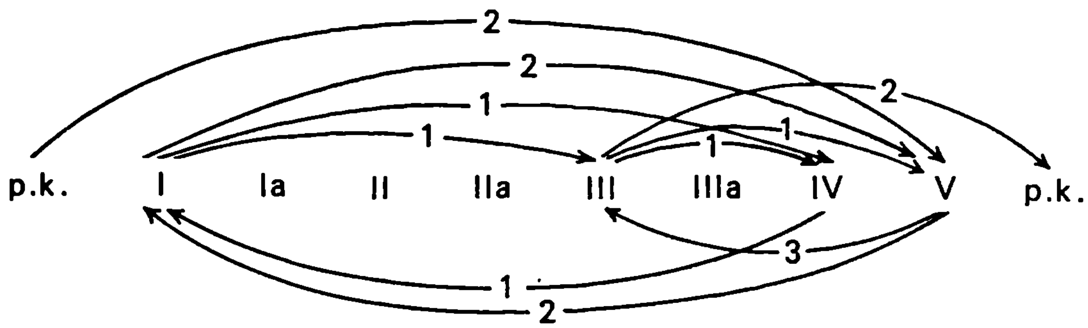
Comparison of the sequences in the Careg-onen Formation at Tandinas Quarry and Flagstaff Quarry (Chart 1) reveals pronounced lateral facies variation. At Tandinas a lower run of dark argillaceous skeletal wackestone/packstones with abundant Daviesiella llangollensis and referable to Lithofacies I are overlain by a sequence of interbedded calcite mudstones, carbonaceous shales and skeletal intraclast, oncoid grainstones. Between 20 and 30 m above the conjectural base of the Formation is a package of fining upwards rhythms, each up to 1.5 m thick, which comprise a coarse grainstone base and grade through finer packstones into a capping calcite mudstone phase. In this context the grainstones form an integral part of Lithofacies V. Discrete beds of cross-bedded grainstone are assigned to Lithofacies III and IV. In contrast at Flagstaff Quarry the latter lithofacies predominate. Thick oolitic units in the lower parts of this section appear on thickness considerations to be laterally equivalent to the fining upwards rhythms at Tandinas.

A facies mosaic in which high energy shoals give way laterally to

a)

	p.k.	I	Ia	II	IIa	III	IIIa	IV	V	
p.k.									2	2
I						1		1	2	4
Ia										—
II										—
IIa										—
III	2							1	1	4
IIIa										—
IV		1							2	3
V	2	2				3				7
	4	3	—	—	—	4	—	2	7	20

b)



c)

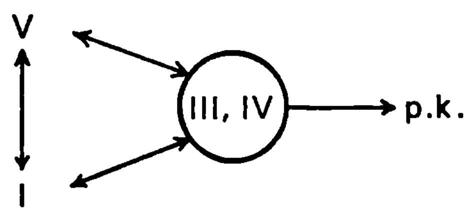


Figure 48

Facies Analysis Diagrams CAREG-ONEN FORMATION

a) Tally Matrix; b) Lithofacies relationship diagram
c) Principal lithofacies associations (see Harms et al, 1975)

peritidal mud flats is indicated, whilst the units of Lithofacies I at the base of the Tandinas section suggest that subtidal conditions prevailed during these lower parts of the Formation. Moreover these units of Lithofacies I are immediately overlain by beds of Lithofacies V without an intervening higher energy shoaling phase. A juxtaposition of a muddy subtidal environment with carbonate mud flats is indicated whilst the monospecific brachiopod fauna suggests somewhat restrictive conditions for the former.

A closely comparable modern setting is encountered off the east coast of the Quatar Peninsula in the Persian Gulf (Shinn, 1973) where chenier beaches of cross bedded grainstone protect tidal flats and sabkhas and build as spits across subtidal embayments. Sediments of the tidal and sabkha flats are typical peritidal carbonates and comprise bioturbated calcite muds with birdseyes, rootlets and stromatolitic lamination. Of particular interest is the occurrence within the low intertidal zone of meandering intertidal channels. These deposit a graded sequence comparable to that of the fining upwards rhythms observed in Tandinas Quarry. The basal, often coarse intraclast oncoid grainstone phase of these rhythms equates with the basal lag developed by these Recent channels with increasingly finer sediment deposited on the flanks of the laterally accreting point bars. The epsilon cross bedding recorded from equivalent deposits within siliciclastic regimes (Reineck and Singh, 1975) has not been recognised in these recent carbonate examples, a function perhaps of the microtidal range, only 1.75 m, of the Quatar coast. Lateral migration of the channels allows for the development of extensive sheet-like deposits in keeping with the geometry of the rhythms at Tandinas.

Alternatively these fining upwards rhythms may record washover events (c.f. Bridges, 1976). The coarser grainstone material then derived from carbonate barrier beaches perhaps represented by the cross-bedded units in Flagstaff Quarry and deposited on the carbonate flats during storms. Comparable spillover lobes have been recorded from the chenier beaches and spits off Quatar.

Sediments accumulating in the subtidal embayments of the Quatar coast comprise highly bioturbated carbonate muds and silts. Spits growing across the mouths of the embayments in response to long shore drift cause increasing restriction and hypersalinity. According to Shinn these subtidal deposits "underlie all the extensive intertidal and supratidal sediments" and therefore exhibit facies relationships comparable with those observed from the Careg-onen Formation at Tandinas. Lower parts of the Careg-onen Formation exposed at Tandinas Quarry therefore record an initial progration and shallowing with somewhat restricted subtidal skeletal wackestone/packstones giving way to peritidal calcite mudstones.

The thick sequences of peritidal carbonates observed in the upper parts of the Careg-onen Formation and indeed at equivalent horizons on the North Wales mainland (George et al, 1976; Somerville, 1979b) have been cited by Ramsbottom (1977) as evidence of a major regressive event at this level the basis for his 5a/5b mesothem boundary, and thought to record an eustatic lowering of sea level. In fact such thick sequences of peritidal deposits reflect sustained deposition at or around base level and are more suggestive of sedimentation keeping pace with a steadily rising sea level. Paradoxically therefore the upper parts of the Careg-onen Formation far from indicating regression may in fact record gradual and prolonged transgression. Nor need this be eustatic since tectonic subsidence combined with an essentially static sea level would generate a similar sequence. The Careg-onen Formation serves well to illustrate the importance of distinguishing simple sedimentary progradational effects from those of true regression i.e. sea level lowering (relative or eustatic) as evidenced by palaeokarst formation.

(b) Facies analysis of minor cycles

Before considering the various lithofacies associations of the higher, minor cyclic formations it is pertinent first to discuss some of the theoretic concepts of cyclic sedimentation.

In their simplest sense the Anglesey minor cyclces record consecutive transgressive and regressive events. Moreover the regressive episodes are truly regressive since palaeokarstic surfaces evidence an actual lowering of base level. Taken as a whole, however, the Dinantian sequence in North Wales is transgressive and records the pulsed inundation of the northern flanks of St. Georges Land (Section 1.4(b)) The overall pattern of base level movements which the Anglesey succession records may therefore be summarised as in Fig. 50a. Schwarzacher (1958) commenting on Dinantian cyclicity in Northern England felt that such repeated "up-and-down movements" of base level were "unlikely". Especially since these events bracket packages of strata of near constant thickness over wide areas and with each transgression effecting an incremental rise in base level relative to an initial notional datum. In fact, given the palaeokarstic framework these "unlikely" parameters appear to have been the reality of carbonate shelf and platform deposition during the Dinantian (Somerville, 1979a, c; Walkden, 1977; Ramsbottom, 1973; Burgess and Mitchell, 1976).

The carbonate sediments overlying Anglesey palaeokarstic surfaces demonstrate the transgressive inundation of these surfaces and the

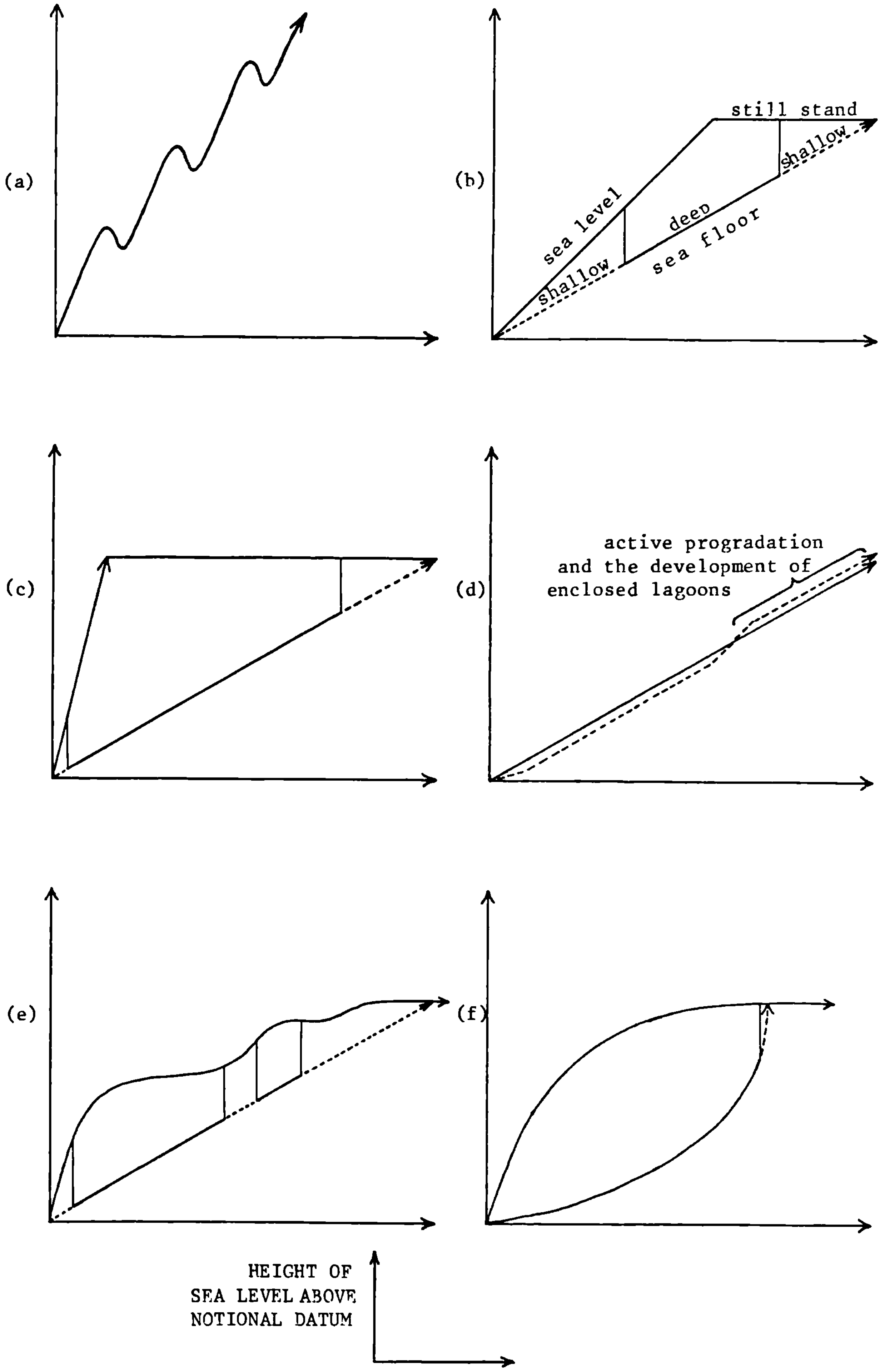


Fig.50 Theoretical interactions between transgressive rises in sea level and carbonate sedimentation.

re-establishment of shelf lagoon conditions (Section 1.4(b)). Capping palaeokarstic surfaces evidence the regressive fall in sea level and exposure of the carbonate shelf. Between these two bracketing events the sediments which comprise the body of the minor cycles were deposited. The better understanding of such cyclic sedimentation rests with an appreciation of how these three factors (transgression, sedimentation and regression) may theoretically combine.

If initially we ignore the effects of regression the sedimentary patterns which form in response to varying transgressive regimes may be discussed. Critical factors are the rates of transgression and of sediment accumulation (upward shoaling and/or progradation). Detailed analysis of the deposits and processes involved in transgressive sedimentation of a Recent carbonate shelf is presented by Logan et al (1969) based on observations of Yucatan Shelf in the southern Gulf of Mexico.

A steady rise in sea level may allow the preservation of units of shallower water, near shore, lithofacies which migrate landward across the shelf with the advancing shoreline. During subsequent periods of still stand these same lithofacies will prograde back across the shelf theoretically forming minor cycles with a symmetrical arrangement of lithofacies (Fig. 50b). With more rapid transgression the preservation of the near shore lithofacies may be precluded and 'deeper water' condition may be established across the shelf almost immediately. These are the 'kick-back' transgressions of Irwin (1965) and characterise carbonate ramps of low gradient where relatively small rises in sea level may inundate large areas of shelf. Subsequent progradation during periods of still stand results in minor cycles with a strongly assymmetric internal sequence of lithofacies (Fig. 50c ; cf. Traeth Bychan Formation). These effects may be further emphasized by the cannibalization of the near shore lithofacies during transgression (Fischer, 1961).

Conversely, if rates of transgression are low then sediment accumulation may be able to keep pace with rising sea level. Deposition of shallow water lithofacies may be sustained with resultant minor cycles composed predominantly of such units. Such aggradation, even progadation during transgression will favour the formation of barrier systems enclosing landward lagoons (c.f. Reineck and Singh, 1975) (Fig.50d). A lateral facies variation from higher energy lithofacies into thick restricted lagoonal and peritidal units may be anticipated (c.f. Caregonen Formation and Moelfre Formation).

These models assume steady rates of transgression, clearly more complex sequences will result where such rates varied or if transgression was pulsed; minor cycles composed of two or more progradational rhythms, perhaps combining the above styles, are to be expected (Fig.50e).

Rates of carbonate sedimentation may also vary. Reduced skeletal production in poorly oxygenated deeper water conditions will result in slower sediment accumulation than in shallower settings where skeletal production is high. Once the latter conditions are established, therefore, rapid shoaling and progradation will result. Thus the linear rise in the level of the sea floor, shown in Figs. 50a to e will more realistically be curved (Fig.50f). The effects of terrigenoclastic influx in damping down carbonate production may also be of importance in this context as may salinity and climatic variation (Wilson, 1975; Flugel, 1982).

The effects of regression do not fundamentally alter these theoretical models, the main result will be to accelerate progradation if shoaling to base level has not already been achieved. A falling sea level will cause the more rapid seaward migration of near shore facies belts than would be achieved by passive sedimentary progradation (Mathews, 1974). Progradational sequences prompted by regression should theoretically, therefore, be characterised by only thin units of shallow water lithofacies. Minor cycles with thin cappings of shallow water lithofacies may, however, also indicate only a shallow wave base reflecting the effects of shelf edge barriers.

The defining palaeokarstic surfaces of the minor cycles strongly suggest that capping shallow water lithofacies were deposited in response to regressive progradation. There are, however, no objective criteria, at least within the geographical limits of the Anglesey succession to distinguish between passive progradational sequences, where regression occurred after shoaling to base level had already been achieved; and progradational sequences induced by regression. If the former situation prevailed there will exist a close correlation between minor cycle thickness and the amount that sea level rose during the initial transgression. If regressive progradation took place minor cycle thicknesses provide only a minimum indication of this transgressive rise.

In this context it should also be remembered that parts of the progradational capping phase will be lost during palaeokarst formation. A factor which may account for the virtual absence of peritidal carbonation of Lithofacies V as an expected veneer to these progradational sequences. Remnants of such units are preserved at the tops of some minor cycles e.g. Porth y Rhos Beds at Traeth Bychan. In other instances the loss of capping units is demonstrated by the occurrence of shoal lithologies as blocks and pebbles within palaeosol regolith, but unrepresented in underlying in situ strata e.g. ooid grainstones in the palaeosol at the top of the Upper Helaeth Beds.

As a final rider to the above discussion it should be borne in mind that these various transgressive/regressive models operated probably not on simple linear facies belts but a complex mosaic of beaches, banks, embayments, shoals and deeps. Facies interpretation however often rests with subjective assessment requiring "a conceptual framework which is a simplified, static presentation of what was probably a dynamic system of inter-related environmental factors operating with varying intensity in different places and at different times" (Logan et al, 1969 p.84).

(c) Flagstaff Formation (Charts 3 and 4)

The facies analysis diagrams (Fig. 51) for the minor cycles of the Flagstaff Formation (average thickness 6.8 m) reveal few recurrent patterns for the formation as a whole; taken singly or in groups of two or three, however, more informative facies relationships are discerned.

(i) Lower groups of cycles (Fl to F3, Lligwy to Forllwyd Beds)

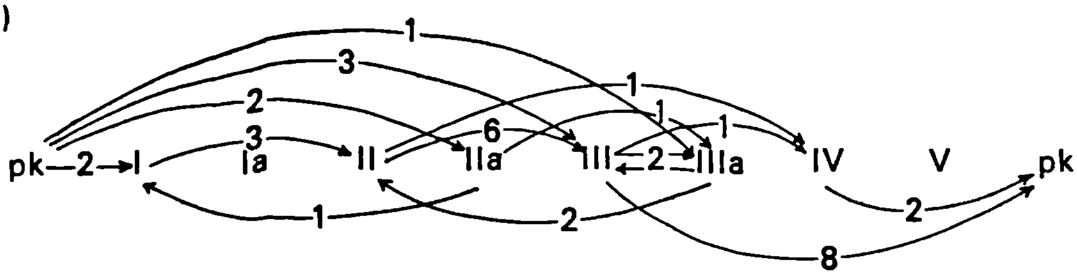
In the Penmon Area these cycles exhibit an intercalation of Lithofacies III and IIIa, the former occurring both at the base and top of the cycles which are therefore broadly symmetrical in their lithofacies distribution. The basal grainstone phases are generally bioturbated. Overlying units of Lithofacies IIIa contain thick (up to 45 cm) locally cross bedded and laterally extensive grainstone beds. The capping, often dasycladacean rich, grainstone phases of Lithofacies III rest with sharp, possibly erosive, contact on the underlying lithofacies and exhibit variable cross bedding directions (Fig. 52).

The basal bioturbated grainstone phases are thought to record the passage of high energy carbonate beaches or bars during initial transgression, with subsequent colonisation of the residual deposits

a)

	pk	I	Ia	II	IIa	III	IIIa	IV	V	
pk		2			2	3	1			8
I				3						3
Ia										–
II						6		1		7
IIa		1					1			2
III	8						2	1		11
IIIa				2		2				4
IV	2									2
V										–
	10	3	–	6	2	1	4	2	–	38

b)



c)

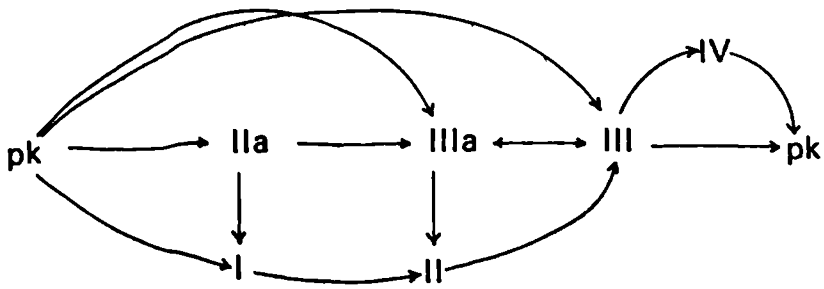


Fig. 51 *FACIES Analysis Diagrams : FLAGSTAFF FORMATION*

a) Tally Matrix; b) *Lithofacies relationship diagram*

c) Principal lithofacies associations (see Harms et al 1975) by burrowing benthos (c.f. subfacies IIa). The upper phases of these cycles suggest the progradation of beaches and/or the lateral accretion of bar or spit crest lithologies (Lithofacies III) across their deeper toe or flank deposits (subfacies IIIa).

Analogous sequences have been recorded from exposed coastlines in the Persian Gulf (Purser and Evans, 1973) where storm beaches are prograding across adjacent intertidal and subtidal deposits. The latter are bioturbated in their lower parts but give way upwards to higher subtidal and intertidal cross-laminated skeletal sands with intercalated thicker cross-bedded units. The details and origins of these features are not discussed by Purser and Evans but on such exposed coastlines it seems likely that they will reflect the effects of constant wave and intermittent storm activity. The capping beach sands of these Recent sequences are locally oolitic or rich in grains of calcareous algae the latter derived from off-shore areas of skeletal production. Cross-bedding is directed on and off-shore whilst the greater spread of vectors obtained from the Flagstaff Formation may indicate additional longshore components.

The Lligwy and Forllwyd Beds of the Principal Area differ from their equivalents in Penmon in exhibiting intercalated siliciclastic units, the Lligwy Bay and Forllwyd Sandstones (Chapter 4, Section 7b) These pass laterally into the main body of the basal Lligwy Sandstone (Fig. 11). Facies description and interpretation for these sandstone units is provided in Sections 4.7 and 4.8. Upper parts of both units comprise cross-bedded and ripple marked calcareous sandstones (Lithofacies C of the siliciclastic scheme; Section 4.4). The overlying carbonate phases of the minor cycles comprise lower bioturbated skeletal packstone/grainstones of Lithofacies II capped by cross-bedded grainstones rich in dasycladacean algal grains and, in the Lligwy Beds, also oolitic.

The calcareous sandstones record the transgressive reworking of terrigenoclastic material carried onto the carbonate shelf during the previous regressive phase and evidence deposition under beach and upper shoreface conditions (Section 4.5a). The succeeding carbonate parts of these minor cycles were probably deposited following the cessation of transgression and in comparison with Penmon, record the regressive progradation of skeletal and oolitic beaches and shoals.

(ii) Cycle F4 (Moryn Beds)

This cycle attains a thickness of 18 m and is the thickest of the minor cycles in the Flagstaff Formation. In Penmon it exhibits a basal grainstone phase, over 3 m thick at Tandinas Quarry. This is overlain by the conspicuous dark grey shale bed, over 2 m thick, which forms a major marker horizon within the Flagstaff Formation sequence (Section 2.4c). Overlying highly fossiliferous argillaceous wackestone/packstones (Lithofacies I) pass upwards via packstones and packstone/grainstones (Lithofacies II) into erosively based capping units up to 4 m thick of cross-bedded grainstone (Lithofacies III). The equivalent minor cycle in the Principal Area, the Moryn Beds, exhibits an almost identical internal sequence save for the absence of the basal grainstone phase. The lower parts of the Moryn Beds comprise intercalated argillaceous limestones and shales of Lithofacies I.

In comparison with the lower cycles of the Flagstaff Formation described above the basal grainstone member in Penmon may record the transgressive passage of higher energy environments. The absence of such deposits in the Principal Area demonstrates their local cannibalisation by the transgressing surf-zone allowing deeper water units of Lithofacies I to accumulate immediately over the basal palaeokarstic surface (cf. Fischer, 1961). The occurrence of units of Lithofacies I, both in the Principal Area and in Penmon, combined with the thickness of the minor cycle suggest that the latter records a particularly marked transgressive event which established relatively deep water, open marine (as indicated by the diverse macrofauna) conditions over this part of the shelf lagoon. The internal sequence of lithofacies is closely comparable to that observed in many of the minor cycles of the Traeth Bychan Formation (see below) and is thought to record progressive shoaling and/or progradation.

(iii) Higher group of cycles (F5, F6; Pedolau and Royal Charter Beds)

In Penmon these upper cycles display a simple asymmetric arrangement of lithofacies. Thinly bedded bioturbated and cross laminated skeletal grainstones of Lithofacies IIIa form the basal phase, up to 2.5 m thick in cycle F5. Skeletal packstone grainstones of Lithofacies II intervene beneath the capping cross-bedded or cross-laminated grainstone phase of Lithofacies III. In the Principal Area the Pedolau Beds differ from cycle F5 only in the thickness of the constituent phases and in displaying an impersistent basal unit of subfacies IIa. The Royal Charter Beds diverge more markedly from their Penmon equivalent. The basal phase comprises two thin beds of highly fossiliferous argillaceous limestone assigned to Lithofacies I. Over 4 m of highly fossiliferous beds of Lithofacies II follow. The thin capping phase, lost in places due to erosion at the base of the Helaeth Sandstone, comprises oolitic and coquinoid grainstones with oncolitic rinds developed around the larger shell fragments.

These higher cycles are not easily interpreted and exhibit styles transitional between the lower cycles of the Formation and the cycles of the overlying Moelfre Formation.

(d) Moelfre Formation (Charts 4 and 5)

The generally thin minor cycles of the Moelfre Formation (average thickness 4.3 m) display a relatively simple association of lithofacies (Fig. 54). Lithofacies II predominates with some cycles wholly composed of such skeletal packstone/grainstones e.g. Upper Lookout Beds. Where other lithofacies are present they occur in predictable positions relative to this principal phase and include basal units of Lithofacies IIa and capping grainstone beds of Lithofacies III.

In the Penmon Area the formation has only been examined in detail in the cliff sections east of Fedw-fawr and clearly this precludes an assessment of lateral facies variation. In the Principal Area, however, such variation has been noted. The packstone/grainstones which dominate the minor cycles of the coastal crop are partially replaced by calcite mudstones of Lithofacies V as the formation is traced inland (Section 2.4d).

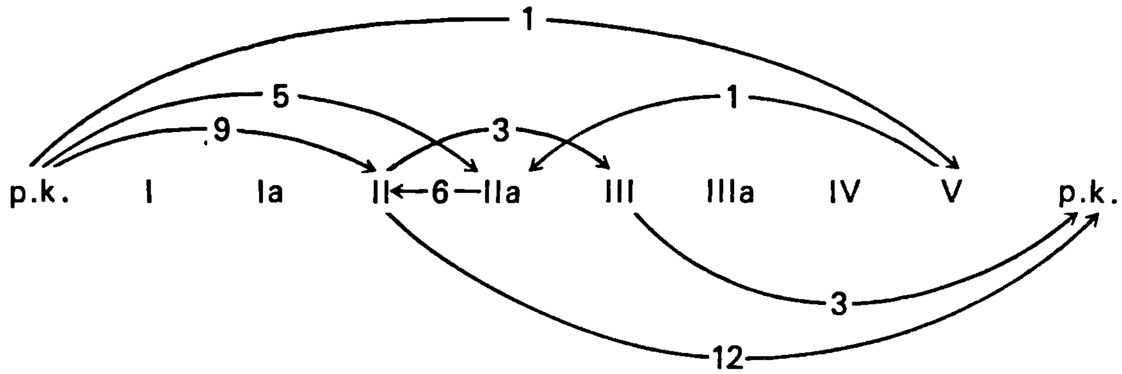
The minor cycles of the coastal sections appear to record relatively simple transgressive/regressive events. In their most complete form i.e. IIa →II→ III they comprise a basal transgressive lag, thick accumulations of subtidal skeletal muddy sands and a capping higher energy shoaling phase. The small thickness of many of the cycles suggests that the transgressive rises in sea level which each records were not pronounced. An interpretation consistent with the absence of mud dominated wackestone/packstone lithologies of Lithofacies I since deeper water conditions were never established, and deposition always took place within the zone of at least moderate water agitation.

The potential to generate such sequences requiring the simple

a)

	p.k.	I	Ia	II	IIa	III	IIIa	IV	V	
p.k.				9	5				1	15
I										—
Ia										—
II	12					3				15
IIa				6						6
III	3									3
IIIa										—
IV										—
V					1					1
	15	—	—	15	6	3	—	—	1	40

b)



c)

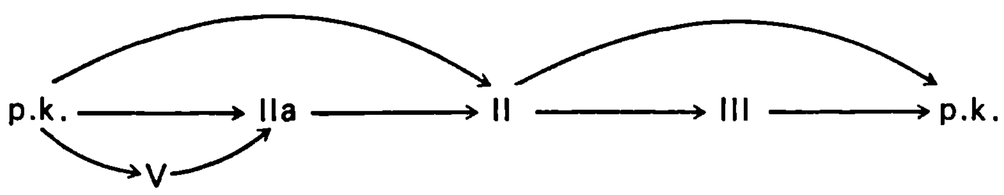


Figure 54

Facies Analysis Diagram MOELFRE FORMATION

a) Tally Matrix; b) Lithofacies relationship diagram;
c) Principal lithofacies associations (see Harms et al, 1975)

progradation of higher energy shallow water sediments across intensely bioturbated subtidal deposits exists in many Recent carbonate settings. A modern analogue of particular interest however has been recorded from Shark Bay, Western Australia by Davies (1970), Read (1973) and Hagan and Logan (1973). Here a three fold sequence of sedimentary units closely comparable with the constituent lithofacies of minor cycles in the Moelfre Formation is described. The sequence comprises a Basal Sheet, a Bank Unit and a Sublittoral Sheet.

The similarities of the Basal Sheet deposits to those of Lithofacies IIa in composition, geometry and position in their respective sedimentary sequences has already been noted (Section 5.5b). Basal Sheet sediments up to 3 m thick floor the deeper parts of Shark Bay (up to 15 m) resting directly on eroded Pleistocene limestones. They contain abundant intraclasts including calcretised grains and lithoskels and are intensely bioturbated. They represent transgressive lag deposits colonised by contemporary benthos.

Sediments of the Bank Unit comprise skeletal packstones and grainstones accumulating in response to the baffling and binding effects of sea grass communities. Individual banks form in a variety of settings and may be fringing, patch, levee or barrier types. Progradation of these banks across the Basal Sheet is an active process and may ultimately generate extensive sheets of sediment comparable with the units of Lithofacies II in the Moelfre Formation.

The textural properties of this ancient lithofacies are mirrored by these Recent deposits. Packstones are developed on the lower slopes of the banks but also in shallow settings where there is a dense sea grass cover. Grainstone textures prevail, generally over the upper portions of the banks, where sea grass cover is sparse and fines are more easily winnowed. Mixing and mottling due to bioturbation occur throughout.

Tidal passes separate adjacent banks and delay their coalescence. The tidal range in Shark Bay is low (up to 120 cms) but the constricting effects of the banks leads to strong tidal currents. The erosively based, intraclast lined grainstone and coquinoid lenses recorded from Lithofacies II may represent the axial deposits of such passes.

As sea grasses die out on the crests of the banks, due to either intermitent exposure or adverse salinity conditions, cross bedded and coquinoid grainstones are developed in response to the now unbridled wave and current activity. These sediments comprise the Sublittoral Sheet and are comparable to the units of Lithofacies III which cap some of the Moelfre Formation cycles. These Recent deposits are similarly rich in peloids, dasycladacean algal grains and grapestone-like aggregates, and are locally oolitic.

In enclosed basins behind the sea grass banks hypersaline conditions prevail and extensive inter- and supratidal carbonate flats are developed. Moreover the sea grass communities only thrive under normal oceanic salinities and the banks therefore only build in a seaward direction. A lateral facies variation from bank dominated lithologies into peritidal carbonates comparable with that observed from the Moelfre Formation in the Principal Area is therefore readily accounted for.

Detailed comparison between the facies patterns of the Moelfre Formation and the Shark Bay sequence is not warranted, however, since sea grass communities only evolved in the Cenozoic (Raup and Stanley, 1971). Read (1974) has suggested the crinoid meadows or thickets of phylloid algae filled a similar niche in the Palaeozoic and may similarly have acted as sediment baffles and binders. It is perhaps also likely, however, that other floral communities which leave no fossil record were in existence, achieving these same effects, in ancient seas.

(e) Traeth Bychan Formation (Charts 6, 7 and 8)

The facies analysis diagrams for the Traeth Bychan Formation (Fig.55) reveal a complex array of relationships embracing the full suite of lithofacies. The complexity of the diagrams is partly a function both of the greater number of minor cycles in the formation and of their on average greater thickness (11.0 m) compared with the underlying formations. Repetitive and instructive patterns are apparent however:

(i) The almost standard occurrence of the argillaceous Lithofacies I and Ia as a thick lower phase; the greater thickness of the Traeth Bychan Formation cycles coinciding with the prevalence of such units. Capping packstone and grainstone phases are comparable in thickness to many minor cycles in the underlying Moelfre and Flagstaff Formations.

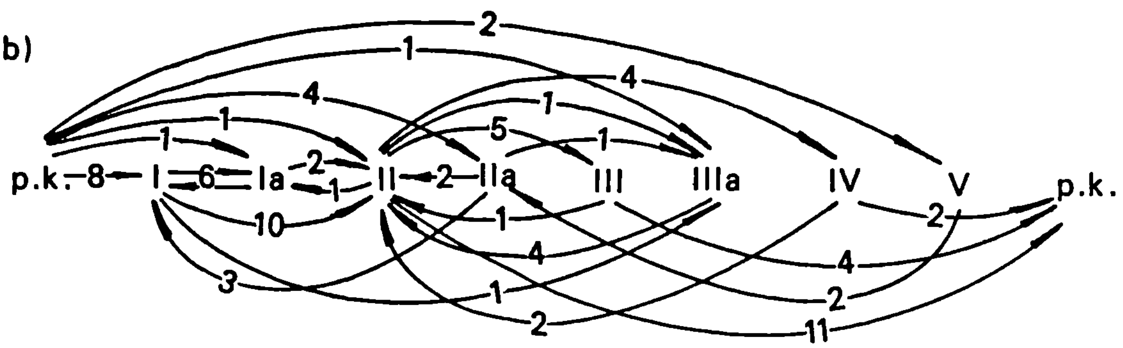
(ii) This lower argillaceous phase often rests directly on underlying palaeokarstic surfaces, or is only separated from them by thin lag deposits of Lithofacies IIa. The lowest parts of these argillaceous strata are often highly fossiliferous but in some of the thicker cycles may give way upwards to more barren runs of the lithofacies which yield only rare productid brachiopods e.g. Port yr Aber Beds and Upper Dinas Beds. A varied suite of trace fossils persists throughout (Fig.42).

(iii) Upper portions of Lithofacies I and Ia are always highly fossiliferous and contain a diverse coral brachiopod fauna. Corals include both solitary and colonial forms, the latter including both fasciculate and cerioid types. These may become so concentrated as form discrete and locally mappable coral beds e.g. in the Upper Dinas Beds at Penrhyn Point. Spinose productids, weighty Gigantoproductids, pedunculate spirifers and thin shelled chonetids dominate the brachiopod fauna. Accessory taxa collected

a)

	p.k.	I	Ia	II	IIa	III	IIIa	IV	V	
p.k.		8	1	1	4		1		2	17
I			6	10			1			17
Ia		6		2						8
II	11		1			5	1	4		22
IIa		3		2			1			6
III	4			1						5
IIIa				4						4
IV	2			2						4
V					2					2
	17	17	8	22	6	5	4	4	2	85

b)



c)

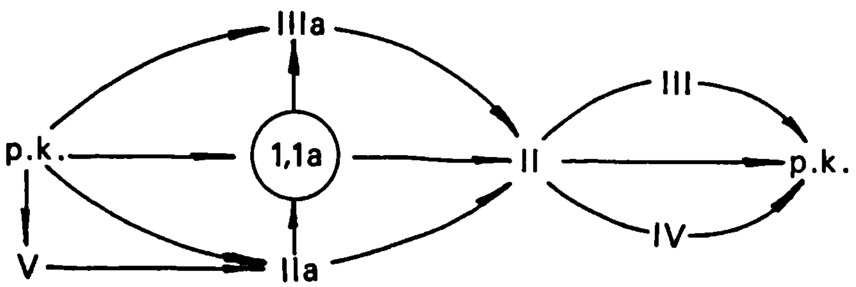


Figure 55
Facies Analysis Diagrams: TRAETH BYCHAN FORMATION

a) Tally Matrix; b) Lithofacies relationship diagram
c) Principal lithofacies associations (see Harms et al, 1975)

or observed within the lithofacies include bryozoans, crinoids, trilobites, hexacorals and othocone nautiloids.

The brachiopods are generally articulated and semi-infaunal types (spinose productids and Gigantoproductids) preserved concave upwards are almost certainly in their life position (Plate 69a). Scattered disarticulated valves are often present however, whilst the tops of some limestone beds may be strewn with disarticulated, concave downward valves, including those of Gigantoproductids.

(iv) These lower argillaceous phases are usually gradationally overlain by units of Lithofacies II although erosively based units of Lithofacies IIIa may intervene e.g. Lower Morcyn Beds. Capping phases of Lithofacies II, III and IV are comparable with the Moelfre Formation minor cycles described previously.

The greater average thickness of the Traeth Bychan minor cycles suggests that they record relatively large transgressive rises in sea level. The pronounced asymmetry of many of the cycles with highly fossiliferous units of Lithofacies I resting directly on palaeokarstic surfaces further indicates the rapid establishment of open shelf lagoon conditions whilst the absence of shallow water, high energy lithofacies suggests cannibalisation of such deposits during transgression (Fischer, 1961). In such instances the underlying palaeokarstic surface is also a plane of surfzone truncation an interpretation strengthened by the occurrence of thin lag deposits assigned to subfacies IIa at the base of some cycles. Lower fossiliferous portions of Lithofacies I may reflect deposition under transitory slightly shallower, better oxygenated conditions than prevailed during the formation of overlying barren sequences of the Lithofacies. These latter units reflect more restrictive environmental conditions and may indicate deposition in somewhat deeper water where slightly anaerobic bottom conditions prevailed, or where dense hypersaline brines were allowed to settle (Purser and Seibold, 1973). It follows from such an interpretation that such units mark the acme of particular transgressive events. Alternatively, unfossiliferous portions may reflect more complete shelf edge barriers preventing circulation and exchange between the open sea and the shelf lagoon. It is worth bearing in mind, however, that a varied suite of trace fossils is often present within these units and that conditions were therefore not entirely injurious to life. Perhaps high turbidity precluded the shelly macrofauna, dominated as it was by suspension feeding taxa.

The return of highly fossiliferous strata in the upper parts of these argillaceous phases indicates the re-establishment of productive environmental conditions possibly through simple sediment accumulation and shoaling but perhaps coincident with the onset of regression. During such upper shoaling phases extensive coral thickets were locally developed.

Layers of disarticulated brachiopod valves within these same strata suggest strong, if infrequent, current action, powerful enough at times to exhume semi-infaunal, thick-shelled Gigantoproductids. These units are comparable with the swell lags of Brenner and Davis (1973) and are thought to indicate winnowing of the mud and concentration of the valves during major storms. Scattered disarticulated valves within beds preserving an essentially in situ fauna are perhaps more likely to reflect the activities of burrowing benthos or of predatory and scavaging necktonic organisms.

Capping phases of the Traeth Bychan minor cycles indicate continued shoaling and record deposition under more agitated conditions at and above wave base such that grain support textures (packstones and grainstones) prevail. They evidence the establishment of shallow water facies mosaics comparable with those developed within the Moelfre Formation (see above), with perhaps vegetated banks of skeletal muddy sand capped by wave worked crest or beach deposits.

Of all the various cycle styles displayed within the Anglesey Dinantian the Traeth Bychan minor cycles compare most closely with shoaling upwards carbonate cycles recorded throughout the geological column (Wilson, 1967, 1975). They are contemporary units to the Yoredale cyclothems developed in Northern England but in contrast are uncontaminated by major siliclastic influx (Burgess and Mitchell, 1976; Ramsbottom, 1973).

(f) Red Wharf Formation (Chart 9)

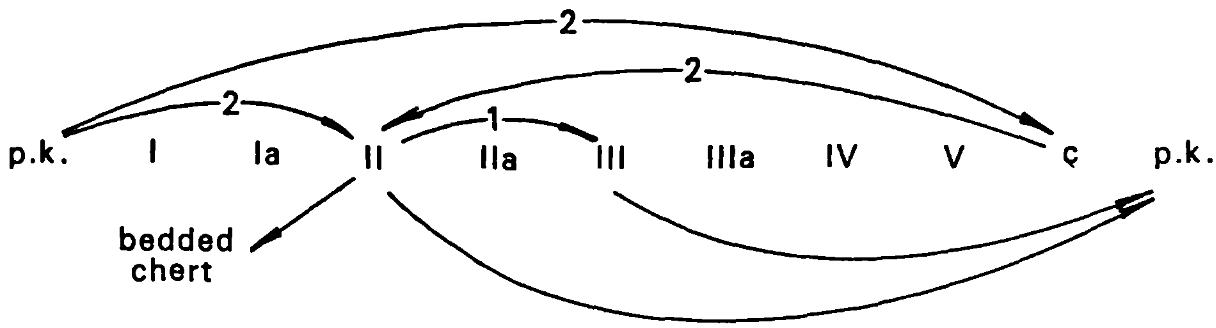
The Red Wharf Formation has only been examined in detail at its type section and is discussed only briefly. It exhibits the greatest average thickness of minor cycles (13.5 m) of any of the formations in the Anglesey Dinantian. Its four minor cycles are composed predominantly of skeletal packstones and packstone/grainstones of Lithofacies II (Fig.56). A minor grainstone phase is developed at the top of the Lower Dwlban Beds whilst the distinctive bedded cherts which cap the Anglesey Dinantian sequence augment the limestones of the Castell-mawr Beds (see section 5.4).

The lithological monotony of the minor cycles precludes their detailed environmental assessment. They contrast markedly with the underlying cycles of the Traeth Bychan Formation for whilst they are of comparable thickness they evidence sustained deposition of relatively shallow water sediments under moderate energy conditions. A response to slow rates of marine transgression and/or high rates of skeletal production , the latter perhaps reflecting the final inundation of the

a)

	p.k.	I	Ia	II	IIa	III	IIIa	IV	V	c	
p.k.				2						2	4
I											—
Ia											—
II	2					1					3
IIa											—
III	1										1
IIIa											—
IV											—
V											—
c				2							2
	3	—	—	4	—	1	—	—	—	2	10

b)



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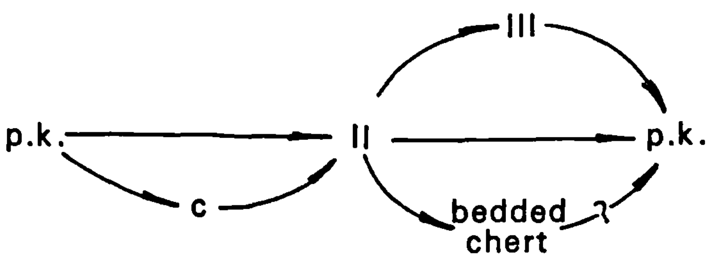


Figure 56

Facies Analysis Diagram: RED WHARF FORMATION

a) Tally Matrix; b) Lithofacies relationship diagram
c) Principal lithofacies associations (see Harms et al, 1975)

adjacent hinterland of older basement rocks and the lack, therefore, of inhibiting terrigenous influx.

CHAPTER SIX

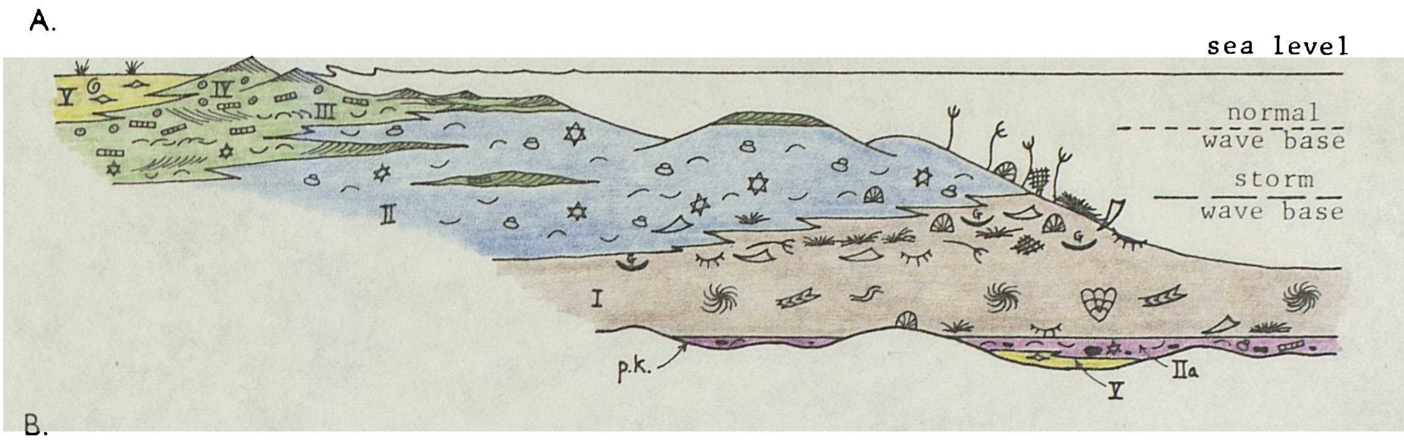
SUMMARY AND CONCLUSIONS

6.1 SUMMARY

The Anglesey Dinantian succession records the establishment and growth of a land-attached carbonate platform. Onlap of underlying basement terrain and marginal terrigenoclastic accumulations is readily demonstrated. Carbonate deposition, however, was repeatedly interrupted, palaeokarstic surfaces recording the periodic lowering of sea level and the emergence of extensive areas of the limestone shelf. The sequence is constructed therefore of numerous transgressive/regressive minor cycles. These have been grouped together into five broader, lithostratigraphically based formations.

During periods of raised sea level active carbonate production was achieved in a mosaic of facies of variable but not pronounced water depth (Fig.57). Calcite mudstone with birdseye structures evidence deposition on peritidal flats protected by grainstone beaches and bars.

Offshore subtidal sediments were predominantly of packstone type and record deposition under conditions of only moderate water agitation or may reflect the baffling and binding effects of benthic or floral communities. Still deeper water environments are indicated by argillaceous but often highly fossiliferous packstones and wackestones, although locally restricted hydraulic regimes allowed such deposits also to accumulate in very shallow settings. Aggradation and progradation of the resulting carbonate lithofacies led to the distinctive minor cycle styles which characterise the various formations.



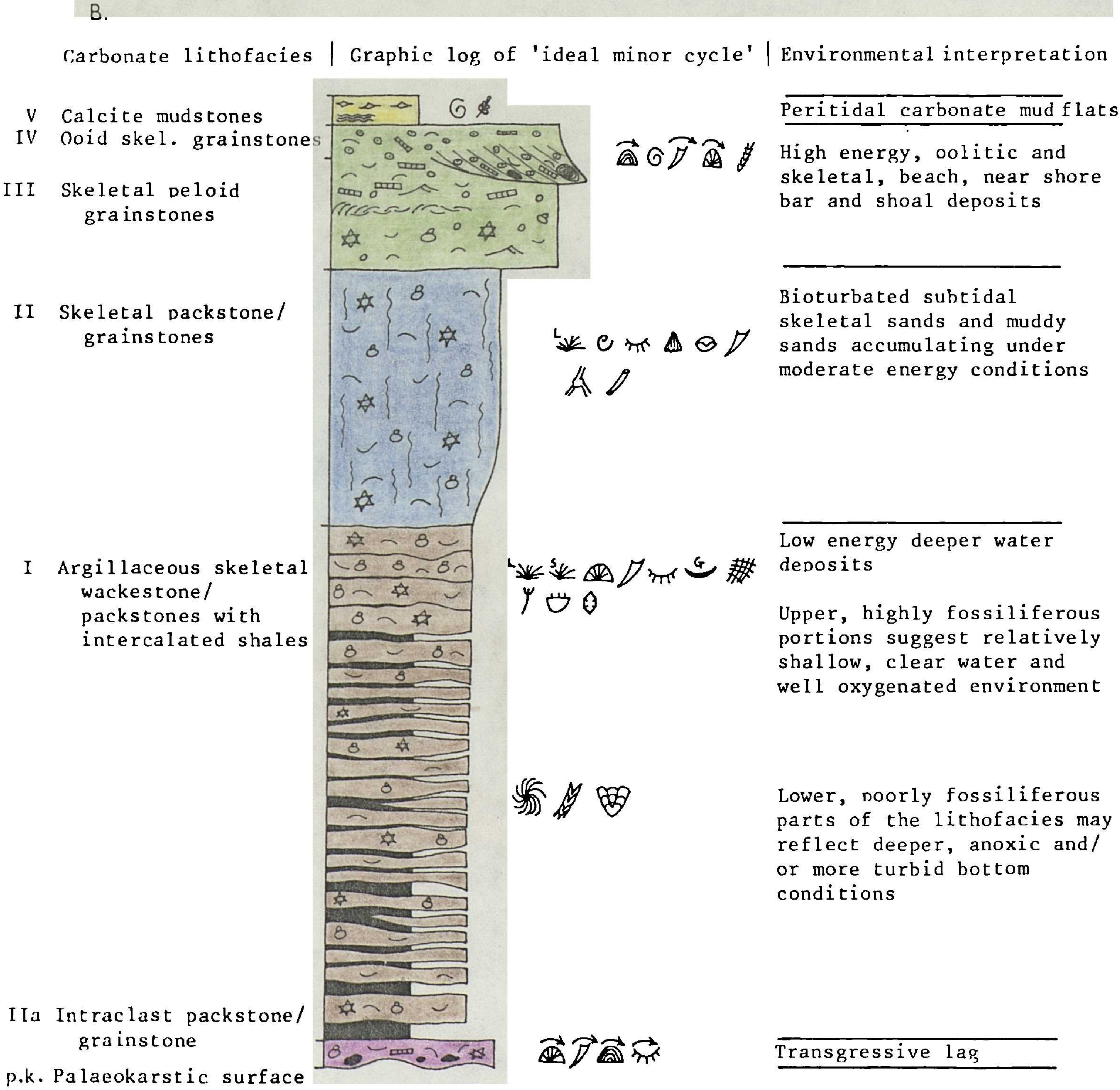


Fig.57 A. Idealized facies belts for the Anglesey Dinantian (c.f. Wilson, 1975); progradation gives the 'ideal' shoaling upwards sequence (B) (see Chart 1 for key to symbols). N.B. upper parts of the sequence, particularly the capping calcite mudstone and oolitic phases are liable to be lost during subsequent karstification (see section 5.6b).

During periods of marine regression and emergence the carbonate platform was subject to the effects of subaerial weathering. Wind-blown volcanic ash accumulated on the exposed limestone surfaces and was colonised and stabilised by vegetation. Complex pedogenic alteration effects were promoted in the underlying carbonate sediment whilst dissolution beneath such soil covers led to the distinctive hummocky topography of palaeokarstic surfaces.

The lowering of erosive base level during regressive periods also resulted in the rejuvenation of siliciclastic source areas within the adjacent hinterland of older rocks. Marginal alluvial fans prograded onto the emergent shelf and rivers incised complex channel systems. Beyond the marginal fans transportation and deposition of terrigenous sediment appears to have largely confined to such channels. The extensive palaeokarstic levels between the channel belts, apart from rare sheet sandstones, preserve no record of terrigeno-clastic deposition. Thus whilst the channel complexes and their contained siliciclastic sequences have been discussed in detail, the emergent limestone shelf was for the most part a zone of terrigeno-clastic by-pass (Galloway and Brown, 1973). The increased volume of siliciclastic sediment generated during regressive episodes was ushered across the shelf within the confines of the channel complexes and presumably debouched into the deeper water, Irish Sea Basin to the north. Palaeokarstic surfaces on the shelf may therefore be expected to correlate with thick packages of coarse terrigenoclastic sediments in the basinal sequence. The verification of such a reciprocal sedimentation model (Wilson, 1967, 1975) should perhaps be the goal of any future stratigraphic exploration beneath the Irish Sea.

Marine transgressions saw the drowning of the channel complexes and the shutting down of terrigenous supply, and culminated in the inundation of palaeokarstic levels and the re-establishment of carbonate facies mosaics on the shelf.

6.2 GROSS TRENDS

The broad distribution of the various carbonate lithofacies present within the Anglesey Dinantian sequence as evidenced by the various formations may be portrayed as in Fig.58. Shallow water lithofacies predominate in the lower part of the sequence, peritidal calcite mudstones passing both vertically and laterally in grainstone shoal units (Careg-onen and Flagstaff Formations). These give way upwards, but are also lateral equivalents of the packstone dominated Moelfre Formation, which in turn is overlain by the thick wackestone/ packstone sequences of the Traeth Bychan Formation. Massive packstones return in the Red Wharf Formation.

Similar gross textural trends have been discerned in carbonate sequences which underlie the modern Bahama Bank (Beach and Ginsburg, 1980) with lower grainstones passing upwards into mud rich packstone and wackestone textured lithologies. The similarity of these sequences to those on Anglesey is strengthened by their cyclic character as indicated by numerous palaeokarstic surfaces. The textural trends identified beneath the Bahama Bank are thought to reflect the evolution of efficient shelf edge barriers leading to increasing restriction and allowing mud grade carbonate to accumulate in increasingly shallower water. Curiously,however an opposite interpretation is favoured for the Anglesey sequence since the thick argillaceous wackestone/packstone units in the higher parts of this sequence (Traeth Bychan Formation) are thought to have been deposited in relatively deep water, whilst their often highly fossiliferous

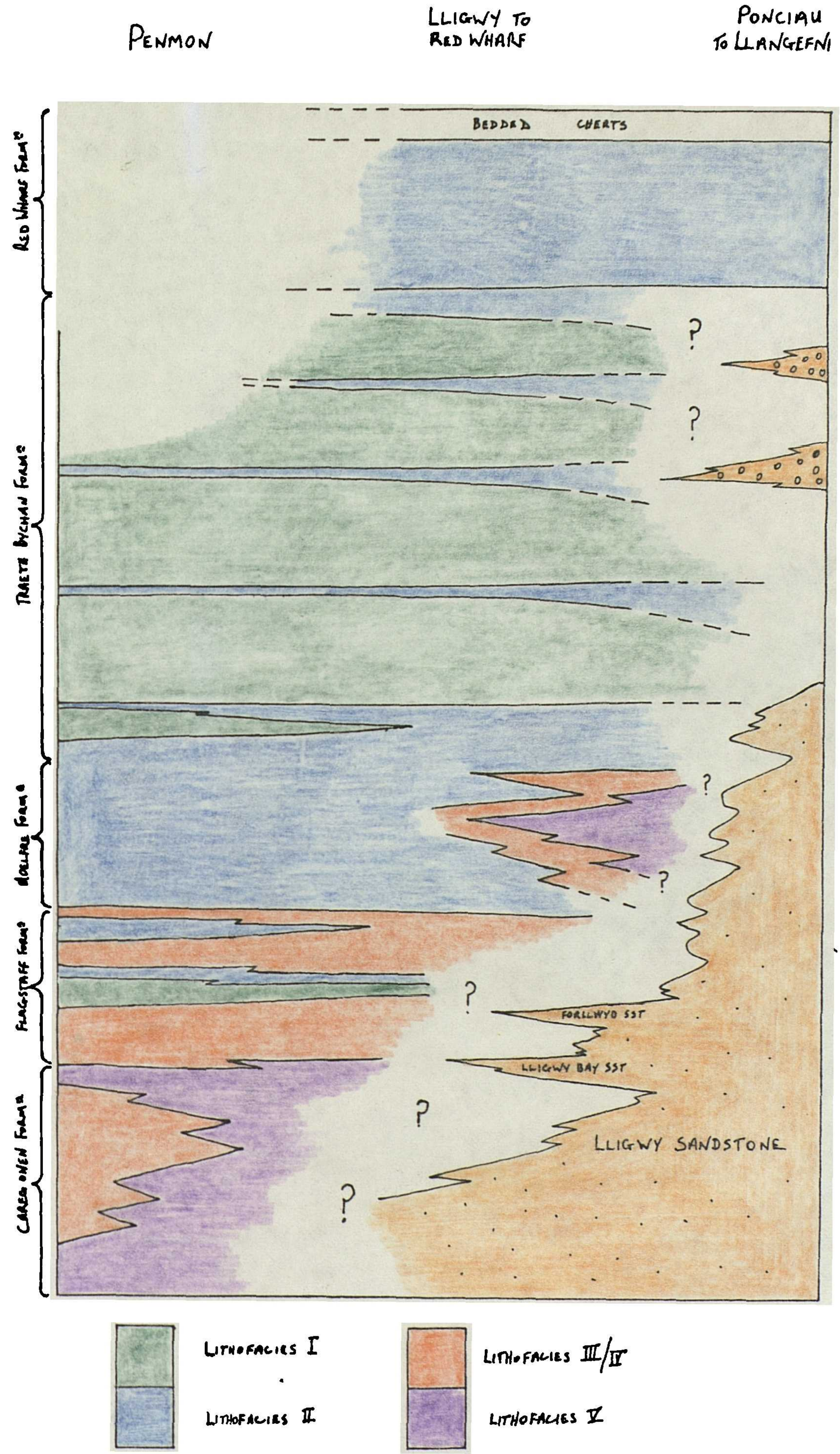


Fig. 58. Gross distribution of carbonate lithofacies

nature suggests deposition under open marine conditions (Section 5.6e). In contrast to the Bahamas, therefore, ineffective shelf edge barriers are indicated, a reflection of the relatively large transgressive rises in sea level envisaged for the thick Traeth Bychan minor cycles and the, perhaps only temporary (cf. Ginsburg and James, 1973) but repeated drowning of high energy shelf edge shoal complexes.

Paradoxically it is the grainstone phases of the Careg-onen and Flagstaff Formations with grain assemblages rich in ooids, pelloids, oncoids and dasycladacean algae that record restricted environmental conditions. Such grains characterise the modern and Pleistocene sediments of the Bahama Bank and indicate depleted skeletal production by normal marine taxa. The prevalence of such restricted grain assemblages, Beach and Ginsburg suggest, is coincident with the onset of more frequent, but less pronounced, oscillations of sea level during the Pleistocene, allowing the development and growth of effective platform edge barriers.

The early grainstone phases on Anglesey are further complicated, however, by the land-attached nature of the Dinantian shelf in North Wales. In this context these lower units occupy a marginal setting. Their restricted character may, therefore, reflect deposition within sheltered coastal embayments (cf. Shark Bay, Logan et al, 1974) as beaches, spits and bars protecting the peritidal, calcite mudstone flats of the Careg-onen and near shore Moelfre Formations.

Both the Anglesey and Bahama Bank successions 'pivot' on thick skeletal packstone sequences, on Anglesey represented by the Moelfre Formation. In the latter, the skeletal grain assemblages indicate derivation from stenohaline shelly communities and in keeping with comparable units underlying the Bahama Bank suggest that open marine conditions prevailed and that shelf edge barriers were incomplete.

These trends, as discussed already, were continued during the deposition of the Traeth Bychan Formation, but the succeeding Red Wharf Formation differs yet again. The thicknesses of the minor cycles in the latter indicate relatively large transgressive rises in sea level whilst the lithofacies composition demonstrates, in contrast with the Traeth Bychan Formation, sustained shallow water deposition. Influencing factors have been discussed above (Section 5.6f). The slower transgressive rises which these units suggests may herald the cessation of the marked oscillations in base level which characterised the Dinantian and the onset of the dominantly progadational styles of cyclicity which predominated during the Upper Carboniferous. Such changes may coincide with the transition across the British Carboniferous cratonic area from 'rift' to 'sag' tectonics (Leeder, 1982; see below).

6.3 CAUSATIVE MECHANISMS

The recognition of palaeokarstic surfaces within the Anglesey Dinantian sequence has been of fundamental importance. These show the succession to be composed of numerous transgressive/regressive minor cycles. It is fitting to terminate this discourse with a brief discussion of mechanisms which may have generated the frequent oscillations of sea level which the minor cycles record. Debate on this subject has focused on the relative importance of two possible contolling factors: eustasy and tectonism (see George, 1978).

The cyclothemic character of upper Dinantian and Namurian Yoredales of Northern England and of Upper Carboniferous cyclicity in general was the subject of an early paper on the subject of controlling mechanisms by Bott and Johnson, 1967. They discuss the various theoretical models and criteria which may be used to distinguish sedimentary cycles which form in response to tectonism from those resulting from eustasy. The assumption is that subsidence of a depositional basin allows the accumulation of a thick sedimentary pile. Distinction between the two possible controlling mechanisms rests with facies relationships within marginal areas where downwarping is reduced. Eustatic movements in sea level will result in the periodic emergence of such marginal sites and the formation of marked erosional discontinuities. Where tectonism alone is active marginal areas are sites of sustained shallower water and nearshore sedimentation, cyclicity is generated by pulsed subsidence of the basin leading to renewed sedimentary outbuilding and the formation of purely progradational rhythms. Bott and Johnson favoured the latter hypothesis for most Upper Carboniferous cyclothems and argue that such subsidence is caused by periodic movements of basement faults in response to the isotatic loading effects of sediment accumulation.

Ramsbottom (1973, 1977) has redressed the balance in favour of eustasy, at least in the Dinantian and Namurian, pointing out the occurrence of widely correlatable regressive phases (see also Section 1.3).

It is important however to appreciate the varying scales of cyclicity described by Ramsbottom. He distinguishes major cycles which 'aproximately coincide' with the regional stages (Fig.2; George et al, 1976) and minor cycles which in the Dinantian appear characteristic of only Asbian and Brigantian strata. The ease of correlation of the major cycles both biostratigraphically and lithostratigraphically makes a eustatic origin easier to accept, but the minor cycles are less easily so correlated and their eustatic interpretation appears to rest on a largely intuitive extrapolation of the major cycle theme (Ramsbottom, 1977).

Ramsbottom (1979) has pointed out that the greater frequency of base level movements evidenced by Namurian cyclothems coincides with the first records of glacial advance in southern Gondwanaland and therefore reflects eustatic oscillations in sea level comparable to those during the last Ice Age. Do Asbian and Brigantian minor cycles reflect eustatic sea level changes resulting from an earlier precursor to this main glacial event? The internal evidence of the Anglesey sequence certainly demonstrates the repeated lowering of marine base level and the formation of erosional diastems (i.e. palaeokarstic surfaces) and therefore, using Bott and Johnson's criteria, argues for an eustatic influence. Yet without detailed correlation of the minor cycles between the various tectonic provinces of the British Dinantian (e.g. Askrigg Block and St. Georges Land) a eustatic control cannot be demonstrated unequivocally.

The blocks and basins which characterised Dinantian palaeogeography in Britain were intra-cratonic in their gross tectonic setting, part of the ancient supercontinent Laurussia (Fig.3; Section 1.4). The onset of minor cyclicity in the Asbian may therefore reflect tectonic pulses which affected the whole of the craton, blocks and basins alike, rather than worldwide eustatic effects. In this context Leeder's work on plate tectonic models for the British Carboniferous is of interest (Leeder, 1982). He suggests that 'hot-creep' within the upper mantle towards the Hercynian plate margin to the south causes tensional rifting in the upper crust generating the blocks and basins of the Dinantian. Attenuation of the crust eventually leads to larger scale subsidence of block and basinal areas alike and to the formation of extensive 'sag' basins during the Upper Carboniferous. The minor cyclic sequences of the Asbian and Brigantian staddle the transition from 'rift' to 'sag' tectonics. During this period, according to Leeder, the margins of the major horst structures (cf. northern flanks

St. Georges Land) undergo uplift as initially hot asthenosphere rises in response to crustal thinning. A purely tectonic mechanism for generating the relative downwards movements in sea level which the minor cycles record is thus provided. The cyclicity itself may reflect the delayed response of ductile asthenosphere in accommodating the stresses released during brittle tensional deformation of the upper crust. The latter causing rapid subsidence followed by slow and gradual uplift. The mechanisms are, as yet, poorly understood but appear to provide a real alternative to the now often blindly assumed eustatic hypothesis.

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