Lower Carboniferous shelf carbonate palaeoenvironments in North Wales

by

DAVID IAN GRAY

Thesis submitted for the degree of

Doctor of Philosophy

at the University of Newcastle Upon Tyne

FEBRUARY 1981

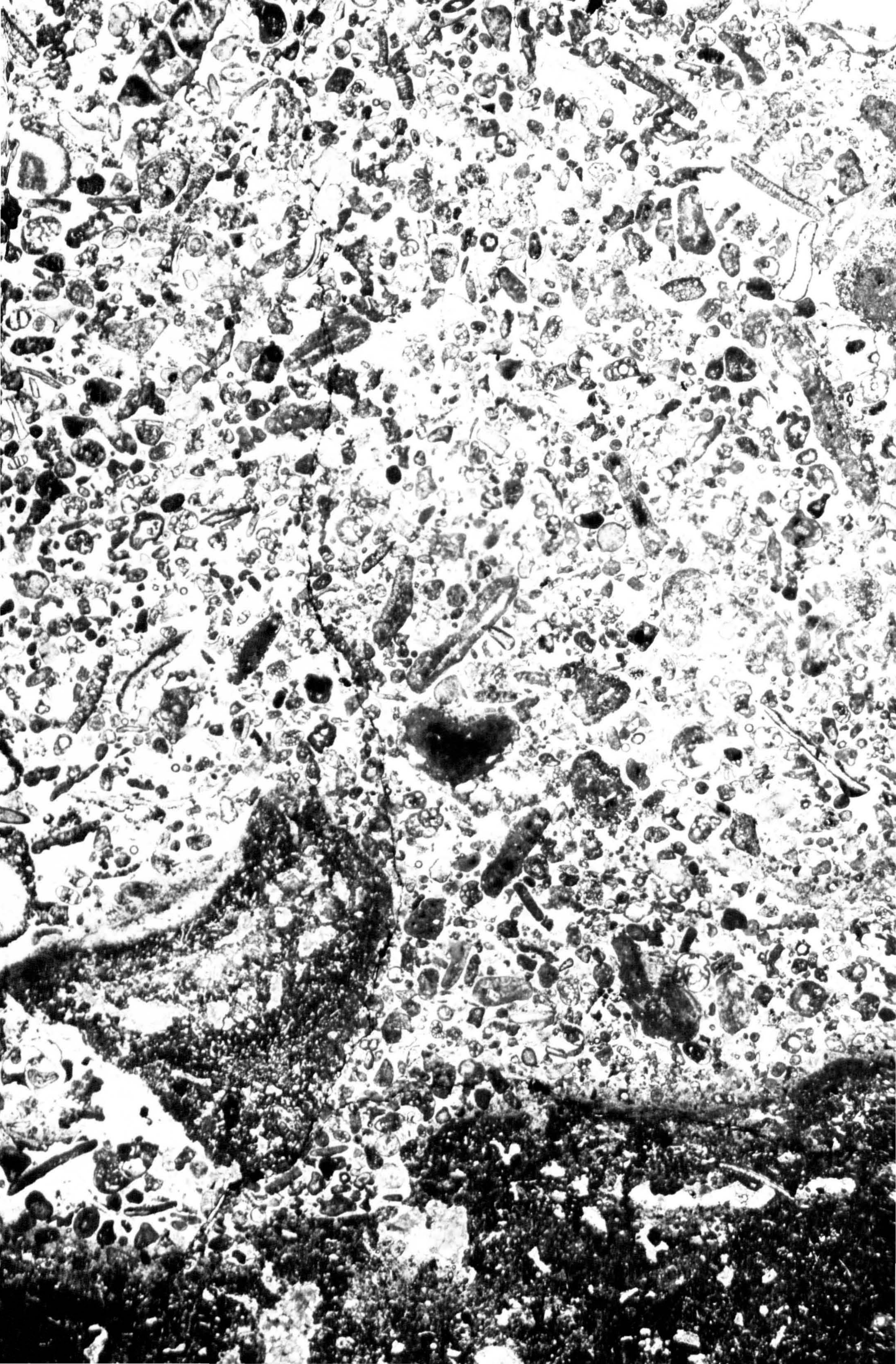
TEXT BOUND INTO

THE SPINE

BEST COPY

AVAILABLE

TEXT IN ORIGINAL IS CLOSE TO THE EDGE OF THE PAGE



Frontispiece

Microscopic character of a minor cycle boundary on the northern flanks of the Llangollen Embayment (Mesothem D5a, Tynant Formation). The erosive top of a fenestrated calcisphere wackestone is overlain by alga peloid grainstone including skeletal oncolites developed around intraclasts, exolithic micrite rims, and mixed generations of bioclasts including lithoskels.

Contact print from negative print from stained acetate peel, x 20.

CONTENTS

List of figures

A B S T R A C T

The Lower Carboniferous (Asbian and Lower Brigantian) of North Wales in the vicinity of Llangollen and Oswestry, records the deposition of about 40 shelf carbonate regressive cycles during a gradual transgression onto the north eastern flanks of the Anglo-Welsh Landmass (St Georges Land). This sequence was deposited during the major eustatic cycles (mesothems) D5a to D6b.

Basement structural elements confined sedimentation during early stages of the transgression to two depositional embayments: a northern Llangollen Embayment and a southern Oswestry Embayment, separated by a tectonic-positive peninsula. Differential subsidence perpetuated these basement influences following their overstep.

A new lithostratigraphy is proposed for the Asbian of the Oswestry Embayment, based on minor cycle character, comprising the Whitehaven (Mesothem D5a) and Llynclys (Mesothem D5b) Formations, and a lithostratigraphy for the lower Brigantian is extended across the region, with revision of existing lithostratigraphy.

Petrographic study of the carbonate lithofacies has revealed 14 carbonate microfacies, grouped into 4 microfacies associations: Calcisphere Wackestone; Argillaceous Alga Packstone; Alga Peloid Grainstone ; Bioclast Peloid Rudstone.

Mesothem D5a comprises about 20 minor cycles, mostly 2-4m thick. Their transgressive phases comprise thin shore-zone conglomeratic lags, overlain by Argillaceous Alga Packstones dominated by beresellid algae. This phase thins to more proximal shelf environments, and passes vertically and laterally into Alga Peloid Grainstones, attesting to a transition from sedimentation below to above a normal wave base. Prominent peritidal regressive phases contain a suite of emergence features including: penecontemporary dolomites; fenestral fabrics; vadose cements; inter-tidal erosion-dissolution surfaces; palaeokarsts, and soil zone alterations.

In the Oswestry Embayment, south of peritidal-regressive minor cycles, shoaling minor cycles were apparently contemporaneous, with regressive-phase oolitic sand blankets that restricted tidal flat progradation.

In contrast minor cycles of Mesothem D5b(late Asbian) (at maximum about 14) are mostly 5-15m thick, and are dominated by Alga Peloid Grainstone regressive phases that indicate sediment reworking and movement near or above a normal wave base. Transgressive phases thin towards tectonic-positive elements , and within the Oswestry Embayment an attenuated sequence reflects a more proximal shelf environment. Minor cycle offlap is marked at the end of Mesothem D5b on the proximal shelf.

Carbonate-dominated minor cycles of the lower Brigantian (of which 8 are laterally persistent), are dominated by transgressive phases that thin towards proximal shelf positions. These sediments support a diverse fauna and include minor growth-baffle coral bioherms. These minor cycles represent deposition in more distal shelf environments to their Asbian counterparts.

Ternary diagrams of bioclast component ratios allow comparison between minor cycles of differing forms and age.

Subaerial emergence phenomena characterise many minor cycle boundaries. Their form is partly dependant upon palaeogeography with palaeokarsts developed more proximal shelf to planar erosion and inter-tidal erosion dissolution surfaces. The boundaries are occasionally multiple, and at times converge, primarily during minor cycle offlap.

The distribution of emergence phenomena suggest a palaeoclimatic cyclicity from arid, calcrete forming periods,to pluvial palaeokarst forming periods during maximum regression.

ACKNOWLEDGEMENTS

I express my sincere gratitude to Dr. C. T. Scrutton and Dr. M. E. Tucker for their supervision, guidance and stimulation of interest during all stages of this study.

I also express my thanks to the many other academic and technical staff of the Department of Geology and Electron Optical Unit (University of Newcastle Upon Tyne) for their help and advice.

My mother deserves special credit for her patience during the arduous task of typing the complete manuscript.

I also thank Roger, John, Mike, Mike, Nicky, Tim, Tony and Trevor, research colleagues, for their continued contributions to the working atmosphere within the department.

This research was financed by a N.E.R.C. research studentship.

Part A

INTRODUCTION

1. INTRODUCTION

This thesis is founded on the sedimentology, sedimentary petrology and palaeontology of Lower Carboniferous (Upper Dinantian) Limestones in the vicinity of Llangollen and Oswestry, North Wales. This transgressive sequence of carbonates lies with angular unconformity upon Lower Palaeozoic strata which formed, during the Lower Carboniferous, the north eastern margin of the Anglo-Welsh Landmass known as St. George's Land.

1.1. PRINCIPAL AIMS AND METHODS OF STUDY.

The principal aim of this study is to establish the depositional palaeoenvironment of the Lower Carboniferous sequence exposed between Minera, Clwyd, (8km WNW of Wrexham) and Llanymynech Hill, Powys (8km SSW of Oswestry, Shropshire).

Subsidiary aims of this study are to:

- establish the vertical and lateral variation of the pervasive cyclicity within the sequence
- investigate probable causal mechanisms for this cyclicity
- erect a lithostratigraphy for sediments of the Oswestry area, based on recent advances in both Carboniferous eustatic cyclicity, and carbonate sedimentology.

This research was conducted primarily by detailed sedimentological logging and facies analysis, sedimentary petrology and comparative palaeontology.

Comprehensive sampling and preparation of acetate peels has provided much of the microscopic and mesoscopic data presented herein. Where appropriate, the scanning electron microscope (S.E.M) with X-ray analysis (E.D.A.X), and X-ray diffractometer (X.R.D) have been used to complement petrographic analysis.

1.2. HISTORY OF PREVIOUS RESEARCH

Ever since the 'ramblings' of Lhwyd in 1699 (Smith,1915), the Carboniferous of North Wales has stimulated research and controversy.

The first work of significance on the Lower Carboniferous of the Llangollen and Oswestry area was completed by G. H. Morton (1870, 1878, 1879) who, in a series of papers, erected a lithostratigraphy for the region, based on dominant colouration of the strata (Table 1). Morton (1886), later extended this stratigraphy into the Vale of Clwyd.

No further advances were made until the zonal stratigraphy of Vaughan (1905) was applied to the sequence by Hind and Stobbs (1906). They (op.cit.) placed Morton's Upper Grey Limestone in Vaughan's "D_2" subzone, and the underlying Lower Grey and Brown Limestone and Middle White Limestone within the "D_1" subzone, using the wealth of palaeontological data collected by Morton.

The second geological survey of the Wrexham area was undertaken between 1910 and 1913 by Wedd Smith and Wills, but the memoir was not published until 1927. Jones (1921) produced a perspicaceous account of the micro-character of the limestone succession. Wedd *et al.* (1927), in the memoir (sheet 121), followed Morton's original lithostratigraphic scheme, but placed the "D_1/D_2" Vaughan-subzone boundary at the base of the Middle White Limestone .

Table 1

HISTORY OF CHRONOSTRATIGRAPHIC/LITHOSTRATIGRAPHIC FRAMEWORK OF LLANGOLLEN & OSWESTRY DINANTIAN

STAGES	MOLD	LLANGOLLEN				OSWESTRY		
	Somerville 1977	Morton 1879 (ON) Hind & Stobbs 1906	Wedd et al 1927	Somerville 1977	THIS STUDY	Morton 1879 (ON) Hind & Stobbs 1906	Wedd et al 1929	THIS STUDY
BRIGANTIAN		CHERTY SHALE L.SDST & CONG SANDY LIMESTONE CEFN Y FEDW SDST	CEFN Y FEDW SDST	SANDY PASSAGE BEDS	SANDY PASSAGE FORM'N	CHERTY SHALE L.SDST & CONG SANDY LIMESTONE CEFN Y FEDW SDST	CEFN Y FEDW SDST SANDY LIMESTONE	SANDY PASSAGE FORM'N
BRIGANTIAN	CEFN MAWR LMST FORM'N	UPPER GREY LIMESTONE	UPPER GREY LIMESTONE MIDDLE WHITE LIMESTONE	TREFOR LIMESTONE FORM'N	TREFOR FORMATION	? UPPER GREY LIMESTONE	? UPPER GREY LIMESTONE	TREFOR FORMATION
ASBIAN U	LOGGERHEADS LMST FORM'N	UPPER WHITE LMST LOWER WHITE LMST	LOWER GREY & BROWN LMST	EGLWYSEG LIMESTONE FORM'N	EGLWYSEG FORM'N	UPPER WHITE LMST LOWER SHALE	LOWER LIMESTONE	LLYNCLYS FORMATION
ASBIAN L ?	LEETE LMST FORM'N	LOWER BROWN LIMESTONE OLD RED SANDSTONE	BASEMENT BEDS	TYNANT LMST FORM'N BASEMENT BEDS	TYNANT FORM'N BASEMENT FORM'N		BASALSHALE	WHITEHAVEN FORM'N BASAL SHALE FORM'N

The Oswestry area survey memoir (sheet 137) (Wedd et al, 1929) revised Morton's (1879) lithostratigraphy, and recognised the lateral equivalence of the Lower Grey and Brown Limestone of Llangollen, with the lower succession of the Oswestry district. (Lower Limestones).

Between 1929 and the 1970's little was published on the Llangollen and Oswestry successions. Neaverson (1929, 1946), in reviewing the faunas and palaeoenvironment of the North Wales Carboniferous, reinstated the "D_1/D_2" boundary of Hind and Stobbs (op.cit.).

Banerjee (1969) and Oldershaw (1969) applied a lithofacies model to the uppermost Lower Carboniferous of the Vale of Clwyd and Halkyn Mountain.

Following the recognition of major cycles of transgression and regression (Ramsbottom 1973), George et al.(1976) erected a new stage terminology for the chronostratigraphy of the area, placing the Lower Brown Limestone and Middle White Limestone within the "Asbian", and the Upper Grey Limestone at the base of the "Brigantian". These two stages were approximated to major cycles (mesothems) D5 and D6 of Ramsbottom. Meanwhile, Power and Somerville (1975) recognised a minor cyclicity within the Middle White Limestone and applied tentative correlations of this cyclicity across the North Wales outcrop.

This cycle concept was further expanded and a new lithostratigraphic scheme erected by Somerville (1977, 1979, a,b,c) for the Llangollen and the Vale of Clwyd areas.

He erected the Tynant Limestone Formation (approximate equivalent to the Lower Brown Limestone) the Eglwyseg Limestone Formation (approximate equivalent to the Middle White Limestone) and the Trefor Limestone Formation (equivalent to the Upper Grey Limestone) (Table 1). This lithostratigraphic division was applied to the succession between Bron Heulog and Minera, north of which he (op.cit 1977) erected a new terminology for comparative sequences within the Mold area of the Vale of Clwyd.

Somerville (1979 a,b,c) recognised the shoaling character of the minor cyclicity and related this to eustatic fluctuations, with subaerial emergence phenomena indicating regression and modification of cycle boundaries.

1.3. GENERAL INTRODUCTION TO THE STUDY AREA.

1.3.1. Outcrop pattern.

The present Lower Carboniferous limestone outcrop between Minera and Llanymynech forms the principal subject material for this thesis.

Figs 1 & 2 show the geographic situation of the outcrop and indicate the important locations along outcrop from which most of the data presented herein has been collected.

The northern limit of the study area is bounded by the Bryneglwys Fault, north of which the outcrop is displaced along the flanks of the Vale of Clwyd. Llanymynech Hill is the southernmost termination of the Lower Carboniferous in North Wales.

The outcrop, forming a long narrow N/S trending belt mostly $\leq$ 1km wide, spans strata from the basal Asbian to the middle Brigantian (stages of George et al. 1976, and approximate equivalents to the Vaughan zones "D_1", and "D_2" respectively of earlier workers).

Table 1 indicates the lithostratigraphic terminology used in this thesis, based in part on Somerville (1979 a,b,c). A new lithostratigraphy is erected for Asbian sediments of the Oswestry area. The lithostratigraphic principles of Holland et al (1978) are followed here.

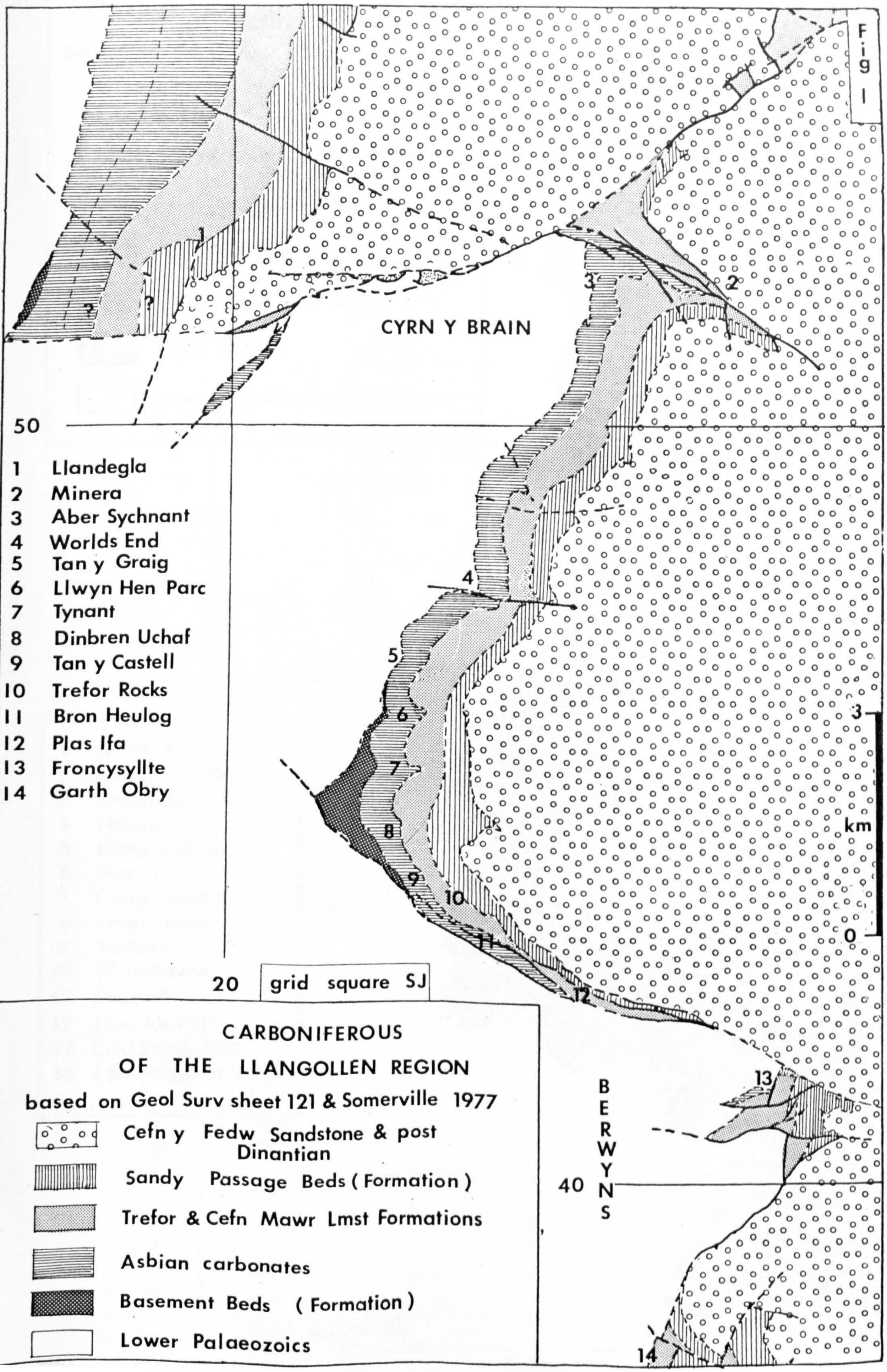
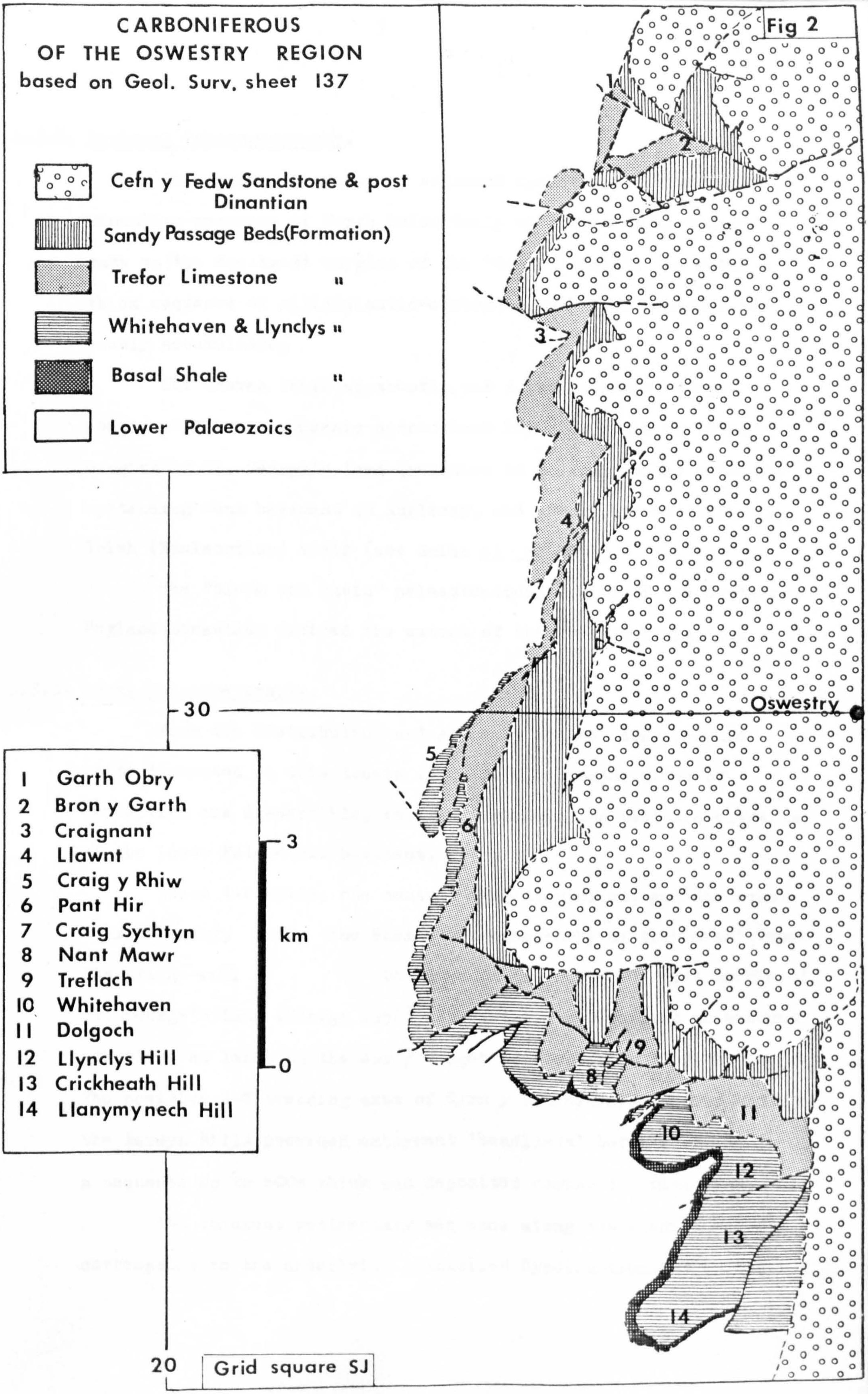


Fig 1
CYRN Y BRAIN
BERWYNS
50
40
20
grid square SJ
3
0
km
1 Llandegla
2 Minera
3 Aber Sychnant
4 Worlds End
5 Tan y Graig
6 Llwyn Hen Parc
7 Tynant
8 Dinbren Uchaf
9 Tan y Castell
10 Trefor Rocks
11 Bron Heulog
12 Plas Ifa
13 Froncysyllte
14 Garth Obry
CARBONIFEROUS
OF THE LLANGOLLEN REGION
based on Geol Surv sheet 121 & Somerville 1977
Cefn y Fedw Sandstone & post Dinantian
Sandy Passage Beds (Formation)
Trefor & Cefn Mawr Lmst Formations
Asbian carbonates
Basement Beds (Formation)
Lower Palaeozoics


CARBONIFEROUS
OF THE OSWESTRY REGION
based on Geol. Surv. sheet 137
Cefn y Fedw Sandstone & post Dinantian
Sandy Passage Beds(Formation)
Trefor Limestone "
Whitehaven & Llynclys "
Basal Shale "
Lower Palaeozoics
1 Garth Obry
2 Bron y Garth
3 Craignant
4 Llawnt
5 Craig y Rhiw
6 Pant Hir
7 Craig Sychtyn
8 Nant Mawr
9 Treflach
10 Whitehaven
11 Dolgoch
12 Llynclys Hill
13 Crickheath Hill
14 Llanymynech Hill
3
km
0
30
Oswestry
20
Grid square SJ
Fig 2

1.3.2. Regional Palaeogeography.

Successive onlapping of sediment cycles that pervade the Dinantian sequence of North Wales imply an accreting carbonate ramp on the southwest margins of the "Craven Basin", in which a thick sequence of siliciclastic-carbonate sediments contemporaneously accumulated.

The Craven Basin, apparently, was a partly enclosed epi-continental sea. Oceanic connections lay around the northern margins of St. George's Land (relative to present north) with the upstanding Mona basement on Anglesey, and across the Southern Irish (Waulsortian) shelf (see Smith et al. 1973).

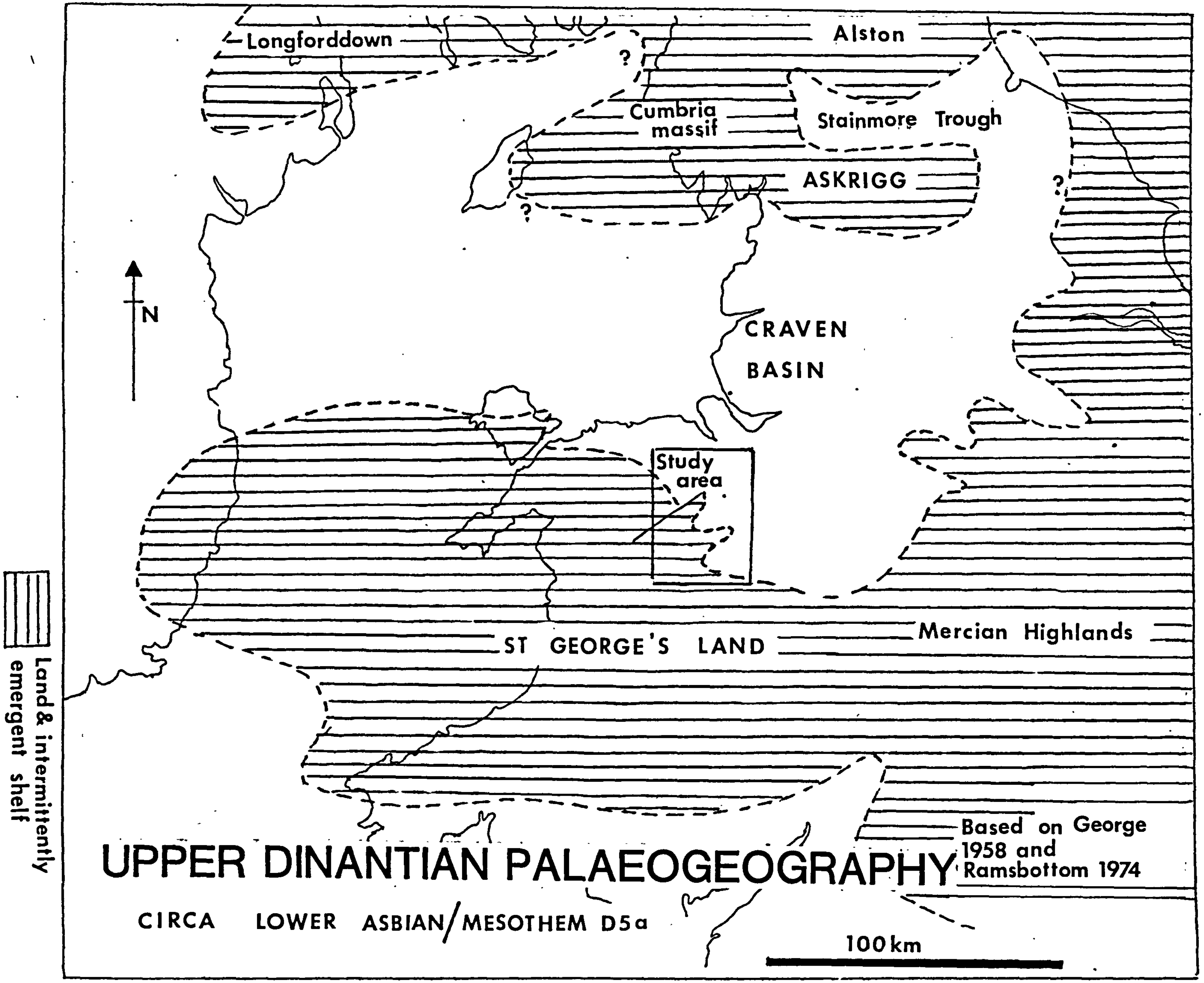
The "block and basin" palaeotectonic regime of the Northern England Dinantian defined the extent of this Craven Basin (fig. 3).

1.3.3. Local Palaeogeography.

From the distribution and geometry of the stratigraphic units discussed in this thesis (fig. 5a), two major areas of deposition are discernible, related to underlying tectonic elements of the Lower Palaeozoic basement.

These two areas, one centred near Llangollen, and the other near Oswestry, show the basic palaeogeography of the 'local' Upper Dinantian well . . Situated on the north and east margin of St. George's Land (George 1958), these two areas reflect an embayed coastline at least in the early stages of deposition (fig. 4). The positive E-W trending axes of Cyrn y Brain/Mynydd Criccor and the Berwyn Hills provided embayment 'headlands' between which a sequence up to 600m thick was deposited during the Dinantian.

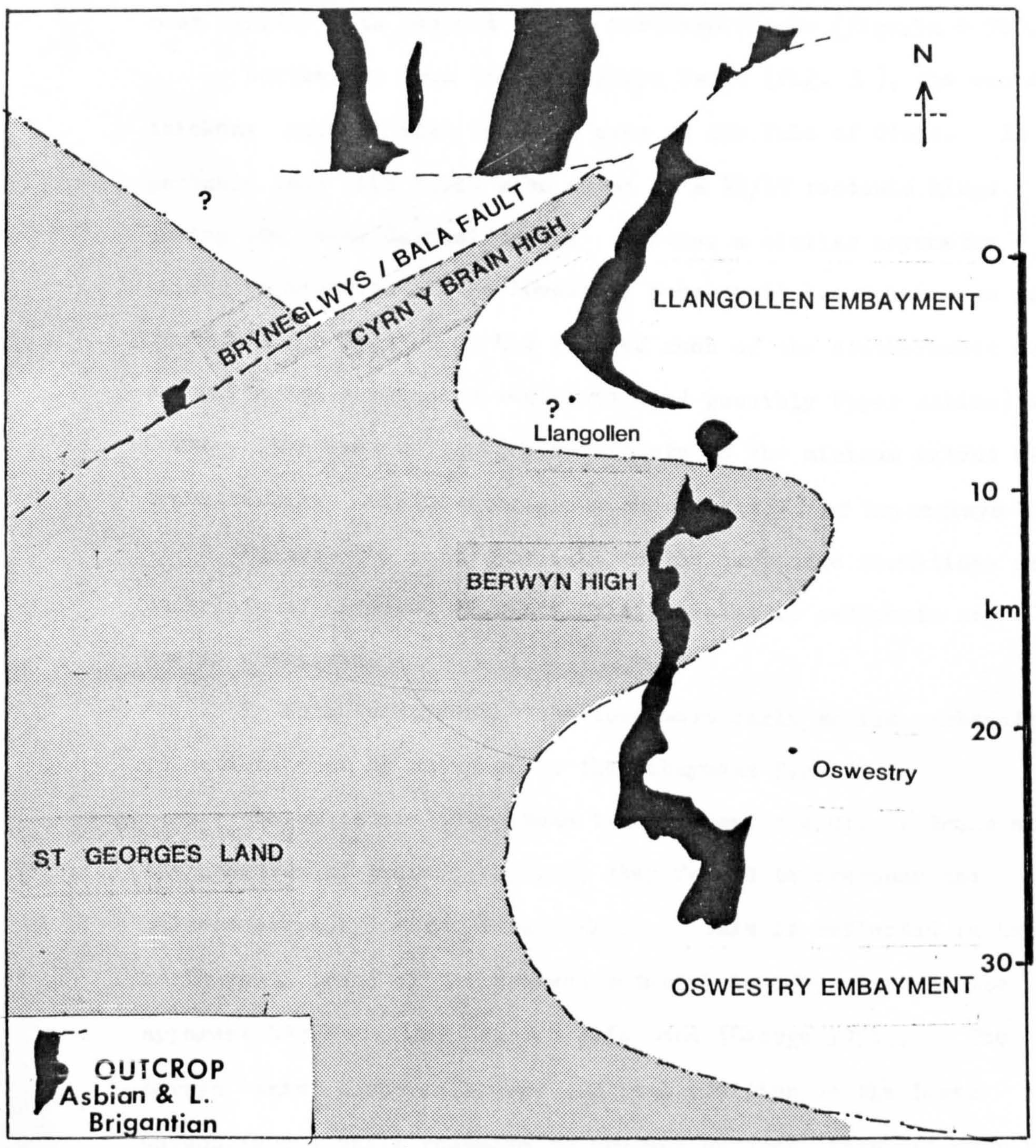
The thickest sedimentary sequence along the studied outcrop corresponds to the underlying Llangollen Synclinorium, which was



Longforddown
Alston
Cumbria massif
Stainmore Trough
ASKRIGG
N
CRAVEN BASIN
Study area
ST GEORGE'S LAND
Mercian Highlands
Based on George 1958 and Ramsbottom 1974
UPPER DINANTIAN PALAEOGEOGRAPHY
CIRCA LOWER ASBIAN/MESOTHEM D5a
100km
Land & intermittently emergent shelf

Fig 3

LOCAL EARLY ASBIAN PALAEOGEOGRAPHY



apparently undergoing the most marked relative subsidence during deposition.

1.3.4. Formation Geometry.

The geometry of 'embayment' sediment fills is asymmetrical. The southern margin of the Llangollen embayment apparently subsided most rapidly with respect to its northern flanks (figs.5a & 5b).

Northwards from the Bryneglwys Fault (fig. 3), the succession thickens rapidly into the Mold area of the Vale of Clwyd. It is probable that this fault also acted as a NE/SW tectonic hinge during the Lower Carboniferous†. Whether a similar asymmetry exists in deposits of the 'Oswestry Embayment' is unclear, as post Carboniferous denudation has removed much of the southernmost outcrop. An outlier of Brigantian sediments (and possibly Upper Asbian) at Corwen, 12km west of Llangollen attests to the minimum extent of transgressing Lower Carboniferous shelf seas (N. of Bryneglwys Fault).

Towards the embayment centres the carbonate formations are underlain by a varied sequence of siliciclastic sediments now mostly unexposed.

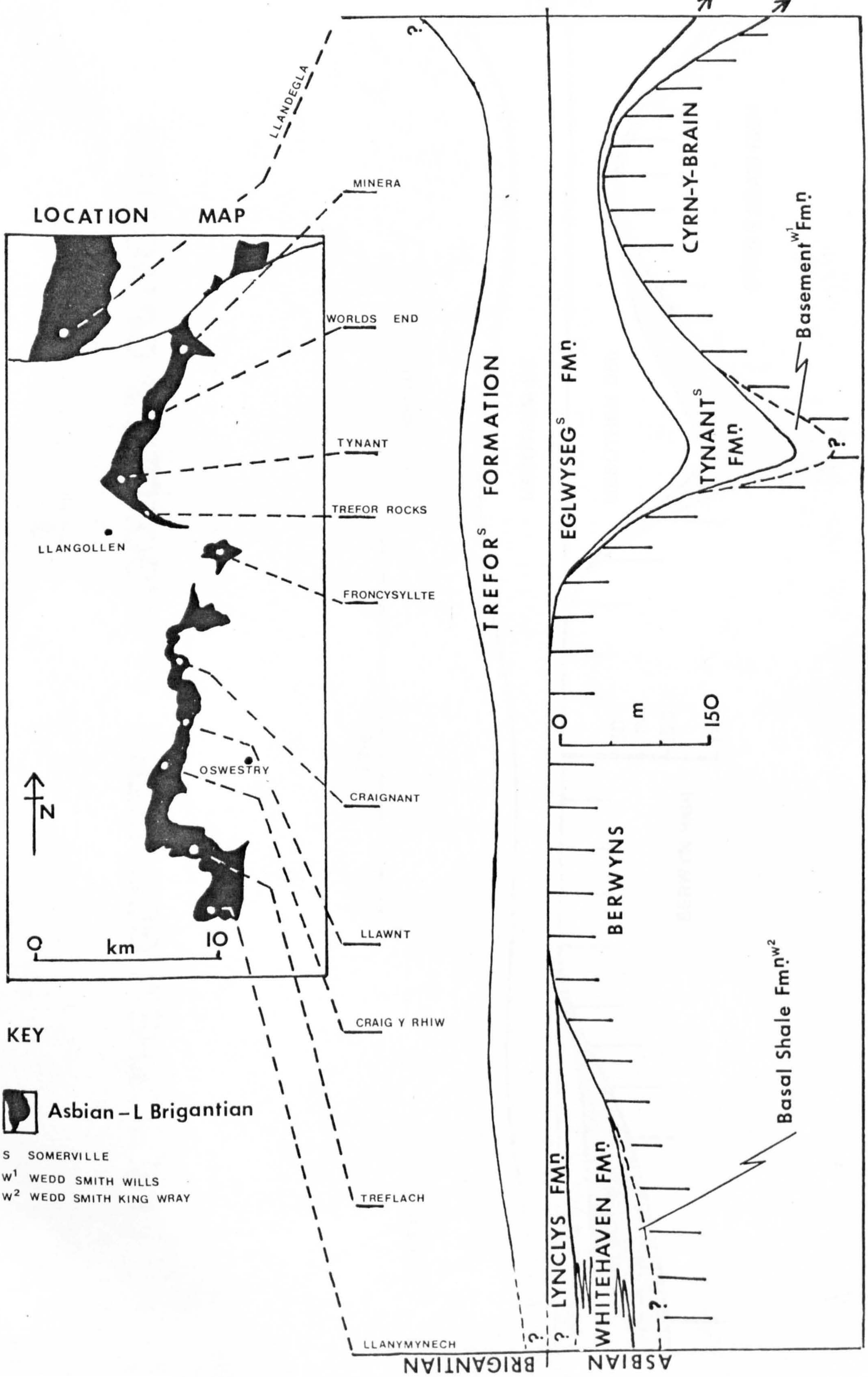
In both 'embayments' the lowermost early Asbian carbonate formations thin by onlap on to the embayment flanks.

Whilst later Asbian seas transgressed the Cyrn y Brain at the position of present outcrop, they failed to overcome the palaeotopography of the Berwyn Hills. This is reflected in the north-south trend of the present outcrop, in comparison to the apparent NW/SE'coast'of St. George's Land (George 1958). The Berwyn 'axis' occupies a more proximal position on the Lower Carboniferous shelf relative to present outcrop, than the Cyrn y Brain 'axis'.

---

† Wedd et al (1927) suggested both pre- and post-Carboniferous movements.

FORMATION GEOMETRY Fig 5a



MESOTHEM GEOMETRY LLANGOLLEN & OSWESTRY

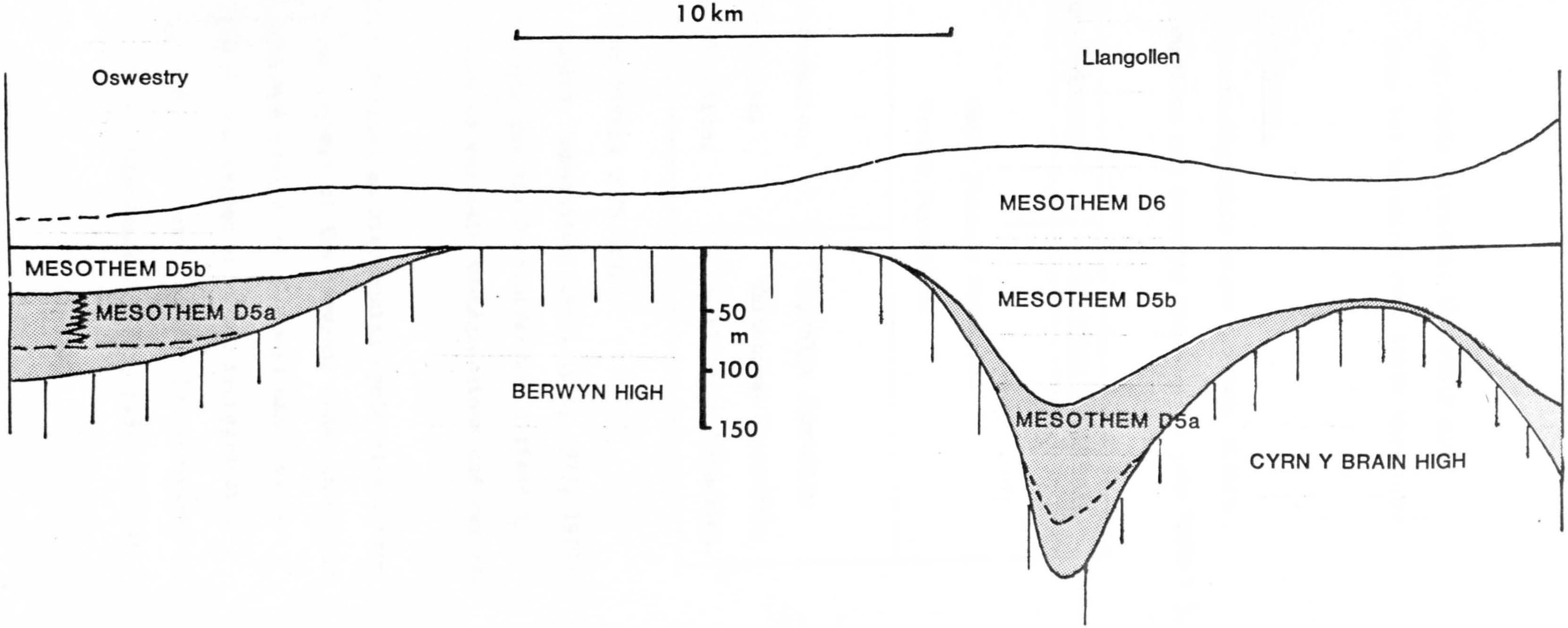


Fig 5b

Lower Brigantian sediments blanketed the study area more uniformly, surmounting, but thinning over these tectonic-positive 'axes'.

1.3.5. Lithostratigraphic Nomenclature.

The following lithostratigraphic terms are used within this thesis for the Llangollen and Oswestry successions (see Table 1).

	Llangollen Embayment	Oswestry Embayment.
Brigantian	Sandy Passage Formation Trefor Formation.	
Asbian.	Eglwyseg Formation. Tynant Formation. Basement Formation.	Llynclys Formation Whitehaven Formation. Basal Shale Formation.

1.4. THE CYCLE CONCEPT AS USED WITHIN THIS STUDY.

In a series of papers, Ramsbottom (1973, 1974, 1977, 1979) erected a cycle stratigraphy for the Carboniferous, defined by lithostratigraphic evidence of eustatic transgressions and regressions.

Ramsbottom (1977) defined an hierarchial cycle classification, recognising, in the context of the sequence under investigation, 'cyclothems' (or minor cycles) as the most minor cycle form, grouped, according to the degree of their transgressive character, into 'mesothems'. This cycle scheme is recognised as a composite 'type B' cycle of Schwarzacher (1975, p.305-306) in the sedimentary succession, and pervades the Asbian and Brigantian of North Wales (Fig.5b indicates the mesothem geometry of the study area).

The term 'minor cycle' is used throughout this study (abbreviated to 'cycle' as this is the basic cyclic unit) in preference to cyclothem. Wanless and Weller (1932) used the term 'cyclothem' to describe sedimentary rhythms from the Pennsylvanian of Illinois. Ramsbottom (in Somerville, 1979a) suggested that the term 'cyclothem' be used for Asbian cyclicity and Duff et. al (1967) stated that "cyclothem" had become interchangeable with the general terms 'cycle' or 'rhythm' in the literature. The cycle form of early Asbian sediments (Mesothem D5a of Ramsbottom, 1977) is markedly dissimilar to that of the Upper Asbian (Mesothem D5b of Ramsbottom, 1977). To avoid a priori assumptions that the term cyclothem represents a particular cycle form, produced by particular cyclic mechanisms (i.e. form and mechanisms comparable to the Pennsylvanian of Illinois), the term 'cyclothem' is not used throughout this thesis. It is recognised that 'cyclothem' may more justifiably be applied to the Upper Asbian and Brigantian minor cycles in the context of previous literature, and the views of Ramsbottom (1977).

Further distinction must be made between 'cyclothem' as a morphological cycle form (Wanless and Weller, 1932, proposed the term 'cyclothem' "to designate a series of beds") or as a genetic cycle form for eustatic cycles (Ramsbottom 1977, 1979).

1.5. THE MICROFACIES CONCEPT AS USED IN THIS STUDY.

The petrographic study of carbonates is fundamental to their description, comparison and interpretation. Essential to these aims is the recognition of groups which allow the investigation of these highly variable sediments.

To help interpret the depositional environments of carbonates, Fairbridge (1954) introduced 'microfacies'. He used both textural and compositional properties of the sediments to define microfacies. Wilson (1975, p.60-69) lists 24 "standard microfacies" which may be encountered on a cosmopolitan scale, each relating to specific palaeoenvironment situations.

There remains, within the literature, a plethora of descriptive formats for carbonate sediments. Part of this variation is due to the increasing tendency to compare past and Recent environments, and to the different aims of many projects.

Lithofacies, genetic facies, microfacies associations and both informal and formally erected microfacies have been used, often in an hierarchial manner, to describe the sediments.

An acceptable descriptive scheme is one which combines field with petrographic observation, i.e. lithofacies with microfacies.

The carbonate terminology used within this thesis follows the classification of Dunham (1962), modified by Embry and Klovan (1969), based on textural relationships within the sediment microfabric.

Within this study, I erect "microfacies associations", defined as a group of fabric-related microfacies.

It follows, but is not implicit within this terminology, that there is a degree of genetic association between microfacies

Table 2

MICROFACIES RECOGNISED WITHIN THIS THESIS

Microfacies.		Petrographic Definition
Calcisphere Wackestone Microfacies Association (MA.1)		
MA.1.1.	Calcisphere mudstone to packstone	p. 74
MA.1.2a	Unlaminated calcisphere-peloid grainstone to packstone	p. 76
MA.1.2b	Laminated calcisphere peloid grainstone-packstone	p. 76
MA.1.3.	Cryptalgal laminite	p. 77
MA.1.4	Laminated beresellid-calcisphere mudstone-packstone	p. 78
MA.1.5.	Coelosporella wackestone to mudstone (Brigantian only)	p. 203
Argillaceous Alga Packstone Microfacies Association (MA.2)		
MA.2.1.	Beresellid wackestone to packstone	p. 124
	or	
	Alga Packstone	p. 199
MA.2.2.	Bioclast packstone	p. 162
MA.2.3.	Calcareous Shale	p. 127
Alga Peloid Grainstone Microfacies Association (MA.3)		
MA.3.1	Alga peloid grainstone	p. 143
MA.3.2.	Bioclast grainstone	p. 146
MA.3.3.	Peloid grainstone	p. 146
MA.3.4.	Oolitic grainstone	p. 164
Bioclast Peloid Rudstone Microfacies Association (MA.4)		
MA.4.1	Bioclast peloid rudstone	p. 153
MA.4.2.	Intraclast rudstone	

Table 3

Lithofacies - Microfacies Association Inter-relationships.

PRINCIPAL LITHOFACIES.	DOMINANT CARBONATE MICROFACIES ASSOCIATION.
Massive bedded pale biosparite.(L.1)	Alga Peloid Grainstone (MA.3)
Thin bedded pale to dark biosparite (L.2)	Alga Peloid Grainstone (MA.3)
Argillaceous thin and or wavy bedded biomicrite (L.3)	Argillaceous Alga Packstone (MA.2)
Stylonodular pale biomicrosparite (L.4)	Argillaceous Alga Packstone (MA.2)
Pale to dark grey porcellanous micrite.(L.5)	Calcisphere Wackestone (MA.1)
Argillite (L.6) :	
Unfossiliferous (Mostly highly coloured (L.6.1)	
Fossiliferous (mostly shades of grey (L.6.2)	
Intraformational Conglomerate (L.7)	
Clast supported (L.7.1)	
Matrix supported (L.7.2)	
Extraformational Conglomerate (L.8)	
Clast supported (L.8.1)	
Matrix supported (L.8.2)	

within a microfacies association.

Each microfacies association has a dominant microfacies after which the association is named. Table 2 lists the microfacies associations and microfacies erected within this thesis.

Somerville (1977) provided a lithofacies analysis of the Llangollen sequence, erecting lithofacies from field and petrographic observations. The lithofacies recognised within this study are given in table 3. Many lithofacies have dominant microfacies associations. Logging and subsequent petrographic analysis has concentrated on the microfacies characters of the sediments, rather than lithofacies. Table 3 indicates the Lithofacies / Microfacies Association inter-relationships recognised in this thesis.

1.6. CYCLE TERMINOLOGY USED IN THIS STUDY

The lithostratigraphic scheme erected here follows mesothem terminology and stratigraphy of Ramsbottom (1977, 1979).

Minor cycles are variable, both vertically and laterally. Upper Asbian and Brigantian minor cycles are readily defined by regression surfaces commonly represented as palaeokarsts, whilst Lower Asbian cycles, of much smaller scale, mostly lack these palaeokarsts and are recognised primarily by regressive phase sediments overlain by those of more transgressive nature.

Within the Upper Asbian, it is the presence of emergence events that leads to recognition of sediment units as individual cycles, especially in more proximal shelf areas (see on). I, therefore, follow here a lithostratigraphic cycle numbering scheme based on these (subaerial emergence) surfaces visible at the type section and within the framework erected by Somerville (1979a,b,) where appropriate:

The following formation abbreviations are used in this nomenclature.

Tynant Formation	Ty
Whitehaven Formation	Wh
Eglwyseg Formation	Eg
Llynolys Formation	Ly
Trefor Formation	Tf

For example, the base of the Eglwyseg Formation is termed

Eg.1 (= top boundary of Tynant Form'n)

The first regression/transgression surface above the Eglwyseg Formation base is termed Eg.2

The intervening 'cycle' is termed

Eg.1.2

The top of the Eglwyseg Formation is Eg.10c = Tf.1

1.7. SEDIMENTOLOGICAL PRINCIPLES OF CYCLIC SHELF CARBONATES

Cyclicity in shelf carbonates and carbonate-clastic sequences is common throughout the geological record. Although the sediment microfacies may change with location and time, cosmopolitan characters attest to similar environments and cycle mechanisms.

Whilst 'transgressive cycles' (e.g. "Lofer cycles" of Fischer 1964) retain evidence of a gradual transgression followed by a rapid regression, 'regressive', or 'shoaling' cycles (e.g. herein) attest to rapid transgression and slow regression.

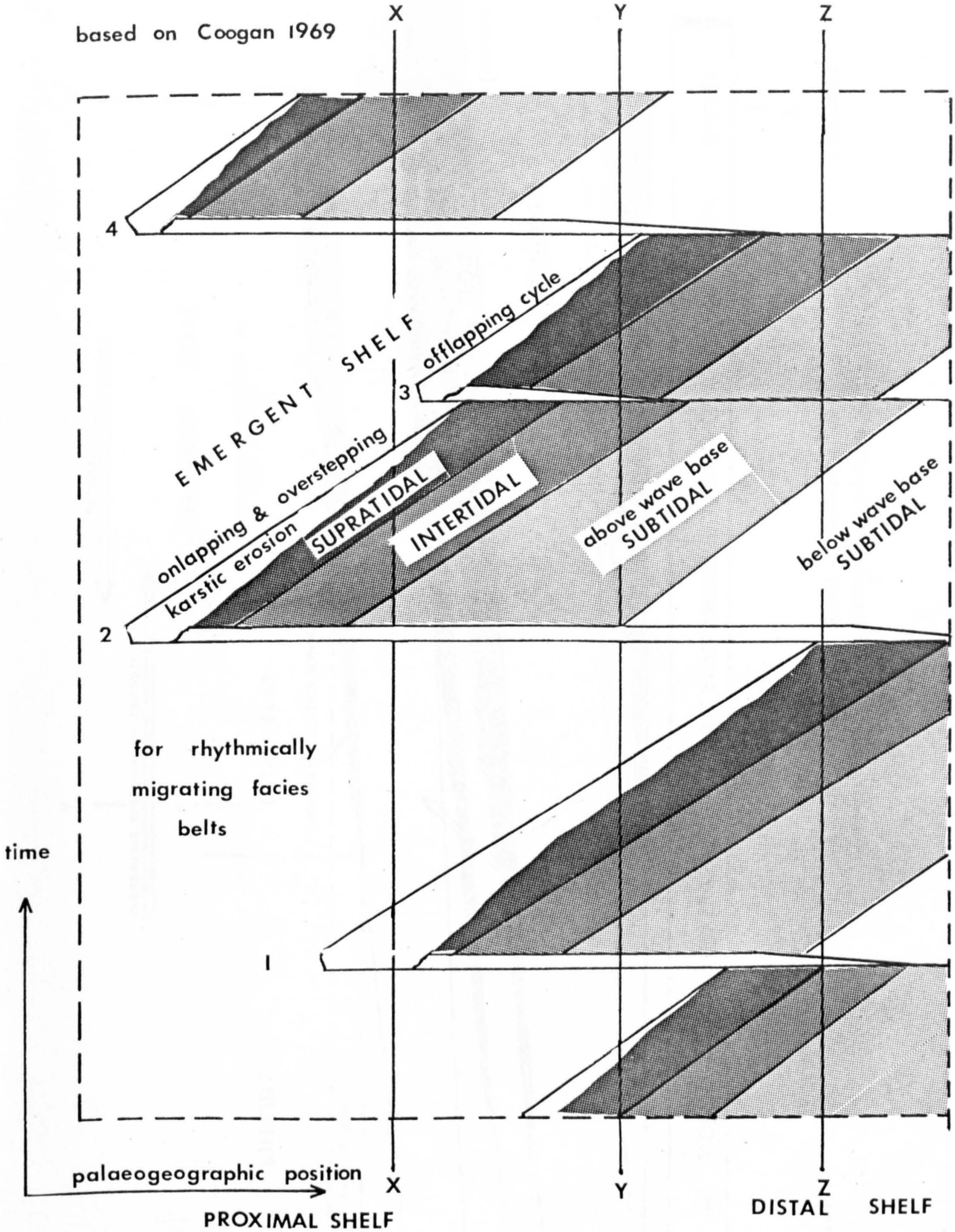
Walker (1979, p.110) depicted a typical regressive cycle model, recognising five phases of deposition from surf zone of the transgressive phase, to terrestrial deposition at maximum regression.

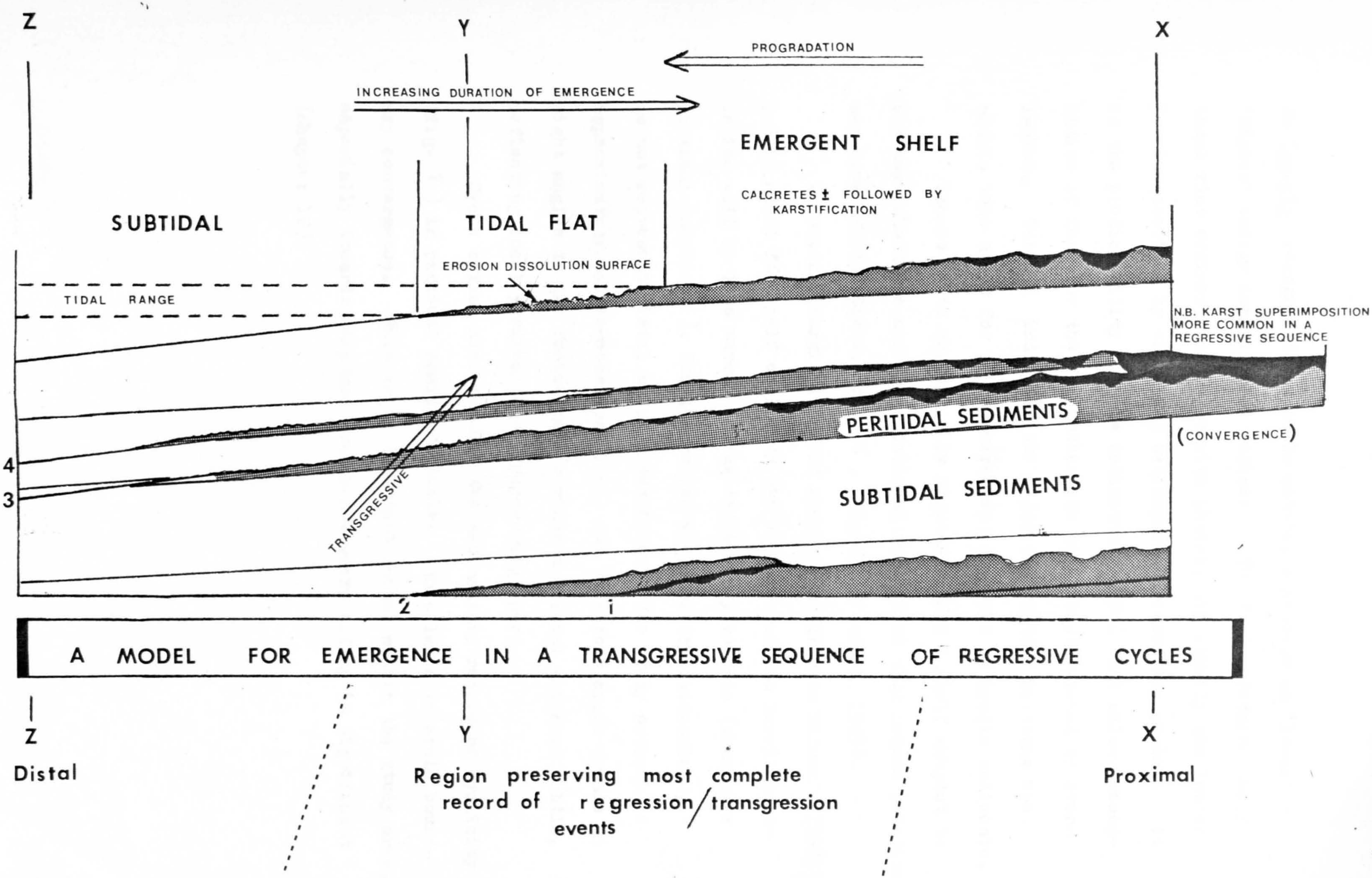
Table 4.

Major sedimentological features with palaeoenvironment significance recognised in this thesis.

Palaeoenvironment	Microfacies & Lithofacies	Sedimentary structures etc.
Supratidal & High Tidal Flat	Intraformational (L.7) & Extraform'nl (L.8) Conglomerates Palaeosol (L.6.1) Calcisphere Wackestone Microfacies Assoc'n (MA.1)	Palaeokarst Calorete structures Rootlet horizons Fenestrae Peds Vadose cements Penecontemporary dolomite
Tidal Flat	Calcisphere Wackestone Microfacies Association (MA.1)	Sutured discontinuity surfaces Stromatolites Pisolites Isopachous cements Impoverished fauna (mainly gastropod) Fenestrae
Shoreface	Extraformational conglomerates (L.8.1) Peloid grainstone (MA.3.3) Bioclast peloid rudstone (MA.4.1)	Cross laminated units Oncolitic grainstones Isopachous cements
Shallow subtidal	Alga peloid grainstone microfacies association (MA.3) Bioclast peloid rudstone (MA.4.1) (L.1 lithofacies)	Megaripple cross-laminated units Ferruginised ooliths Diverse fauna Cross-bed shoals Sediment mixing and algal micritisation Storm layers
Transgressive phase	Argillites (L.6.2) Intraformational Conglomerate (L.7.2) L.2; L.3; and L.4 lithofacies	Diverse fauna *Zoophycos* bioturbation Biostromes and Growth Baffle Bioherms

CHRONO – FACIES MODEL FOR REGRESSIVE CYCLES





A MODEL FOR EMERGENCE IN A TRANSGRESSIVE SEQUENCE OF REGRESSIVE CYCLES

fig 7

He (op.cit) recognised two basic models, dependant on 'lower' or 'higher' energy depositional regimes. The former retains muddy tidal flat sediments as regressive phases, and ideally the latter is characterised by shore zone sediments as regressive phases. It is the predictability of this sedimentary model that allows recognition of features that may otherwise be mis-interpreted or overlooked. Table 4 indicates the major sedimentary criteria used within this study for palaeoenvironment analysis of cyclic sediments.

Facies belts rhythmically migrate across a shelf subject to eustacy subsidence and progradational mechanisms that induce relative sea level fluctuations. (fig 6 , based on Coogan, 1969).

On distal (marginal) shelf areas, according to Wilson's (1975), p.281) model for shelf cyclicity the effects of sea level fluctuation will be less marked in the sedimentary records (than more proximal locations). The Llangollen and Oswestry successions, do not represent distal shelf situations. The study outcrop is approximately palaeo-coast parallel, and therefore cycle trends at right angles to the facies-belt strikes are local and more subtle, reflecting local proximal shelf palaeogeography.

Cycle offlap also reduces the sedimentary record of cyclicity (fig. 7) in proximal shelf locations. This leads to cycle boundary convergence. This is an important factor within the study area, especially towards mesothem boundaries where offlap is significant (chapter 12).

Part B

STRATIGRAPHIC FRAMEWORK

2. STRATIGRAPHY AND CYCLE FORM OF THE TYNANT AND WHITEHAVEN FORMATIONS

The Tynant and Whitehaven Formations are dealt with together because of their similarities in sediment types, cycle structure, and chronostratigraphic equivalence. Both Formations were deposited during Mesothem D5a (Ramsbottom, 1977).

2.1. TYNANT FORMATION

Exposed along the basal part of the Eglwyseg Escarpment, the Tynant Formation thins to the north and south from an outcrop depocentre at Tynant, the type locality erected by Somerville (1977). It is the lowest Asbian carbonate-dominated formation in the Llangollen Embayment and is part equivalent of Morton's Lower Brown Limestone (see Table 1). It is underlain conformably towards the outcrop depocentre by a red-bed siliciclastic 'Basement Formation' ('Basement Beds' of Wedd et al. (1927) and 'Old Red Sandstone' of Morton (1879)) that is no longer exposed. Wedd et al (1927) mentioned early 19th century reports of "dark red sandstone with some brecciated conglomerate interstratified with it", and Morton (1879) records a few exposures of 'Old Red Sandstone' near Tan y Graig, Dinbren Uchaf and Tynant Quarries. None of these are now visible, although its presence is evident from red soils, red fine to coarse sandstone blocks, and reddened conglomeratic sediments. The Basement Formation may be up to 60m thick at Tynant, but thins rapidly north and south from this to zero<4km north and south of Tynant.

The Tynant Formation is overlain, along the whole outcrop by the Eglwyseg Formation. Somerville (1977, 1979b) defined the upper boundary of the Tynant (Limestone) Formation as the junction between the uppermost, thickest and laterally persistent 'porcellanous limestone' with well bedded bioclastic limestones typical of the overlying Eglwyseg Formation, and specifically the basal phase of his Eglwyseg Formation Cycle 1. He also noted the top of the Tynant Formation as the uppermost occurrence of Daviesiella llangollensis.

Somerville divided the Tynant Formation at Tynant into a lower sequence of thin cycles with argillaceous biomicrite and shale basal lithologies,and an upper series of seven variable, but generally thicker cycles, the upper three of which have "thickly bedded non-argillaceous bioclastic" grainstones (Somerville 1979b). At its thinnest northward development in the Minera region (where it is ~ 4m thick) it rests with angular unconformity on the underlying Lower Palaeozoic litharenites. In the Tynant area it is~80-100m thick (120m from data of Morton (1879)) but only the top 55m is exposed. The top 10m oversteps the Berwyns as far as Froncysyllte(Fig. 8.)

Cycle correlation, especially in lower exposed strata is difficult due to the similarity of each successive cycle and the known lateral variations in bed thicknesses, bedforms (especially cycle tops (see chapter 6.)) and microfacies (especially MA.1 microfacies). Correlation can, therefore, be established only with a high degree of confidence in adjacent sections (see Chart A), hundreds of metres apart. Somerville (1977, 1979b) recognised some of the pitfalls of cycle correlation. His correlation chart (1979b, p.398) indicates a number of problems requiring clarification:

1). Correlation of the strata from Somerville's type sequence at Tynant is highly subjective. He used plant bearing and coal horizons (cycle 10 top of Somerville (1979b)), occurrence of 'porcellanous' limestones, and general cycle character as markers. He

MESOTHEM D5a LLANGOLLEN EMBAYMENT

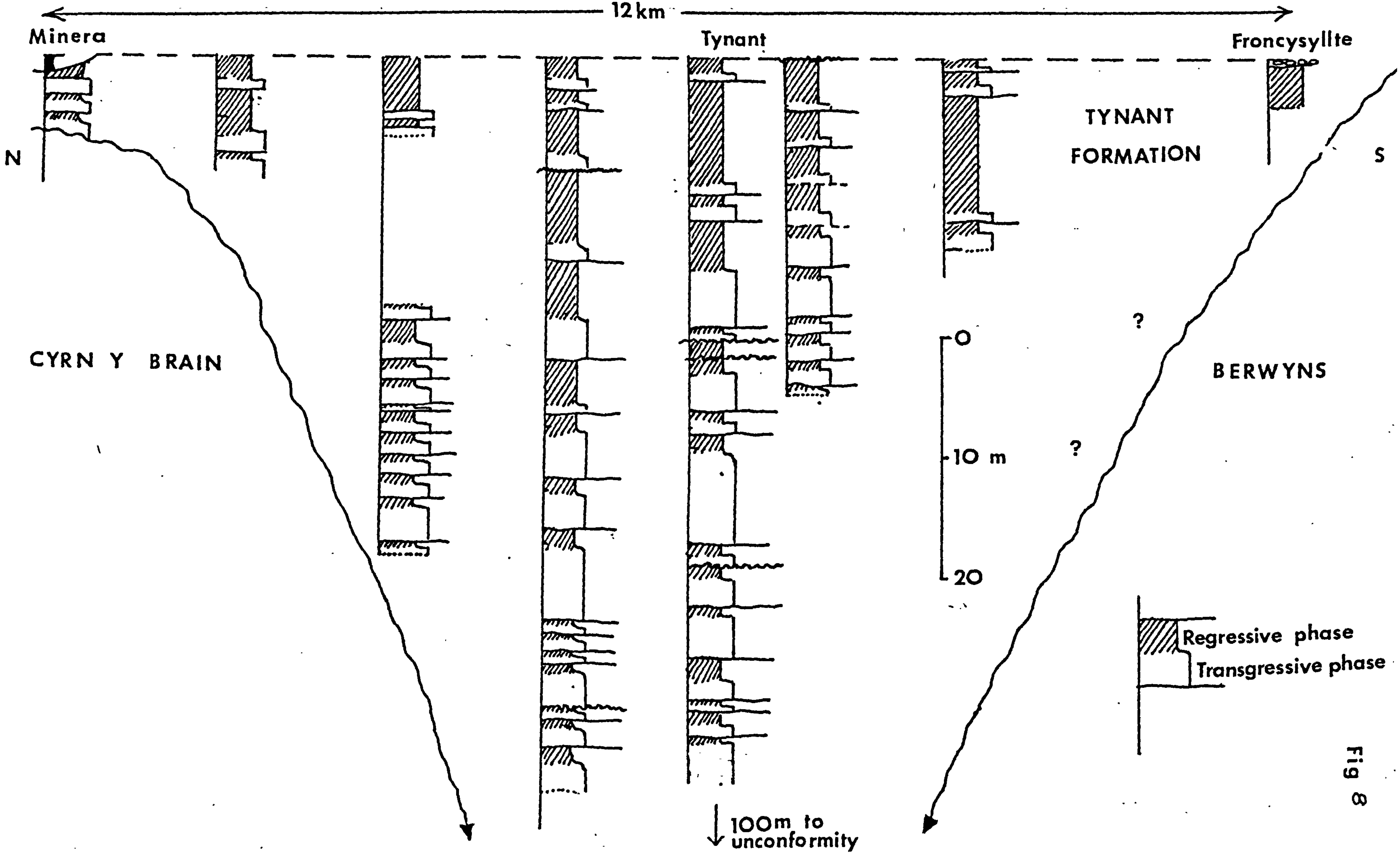


Fig 8

recognised that the 'porcellanous limestones' have limited lateral persistence, and reflected this in his tentative correlations (Somerville, 1979b, fig.1).

2) Somerville recognised 15 cycles in the Tynant Formation as exposed at Tynant, each one defined by an upper phase of Calcisphere Wackestone (MA.1). Detailed analysis of the uppermost cycle and others indicate that some may be considered multicyclic (see figs. 9a & 9b).

3) Somerville noted that the large chonetid Daviesiella llangollensis was 'not known from the overlying Eglwyseg Formation' (Somerville, 1979b, p.397). Chart B attests to its infrequent presence in strata of the first ?and higher Eglwyseg Formation cycles, all along outcrop from Minera to Trefor Rocks in the south.

2.1.1. Cycle Form

In the Tynant Formation the cycles vary between about .50cm and 12m thick. Most are between 1 and 4m (see chart A). Ideally they comprise a basal bioclastic phase of argillaceous alga packstones (MA.2 microfacies association) and an upper phase of calcisphere wackestones (MA.1 microfacies associations) corresponding to the 'porcellanous limestones' and 'calcite mudstones' of previous workers. Intermediate in position within the cycle, thin to medium bedded alga peloid grainstones (MA.3 microfacies association) are variably developed, (figs 11 & 12) but develop as thicker units in cycles towards the top of the Tynant Formation, reflecting a gradual transition to cycles of Eglwyseg Formation character.

Microfacies commonly grade into each other. Minor lithofacies including intraformational and extraformational conglomerates occur in basal phases of cycles, and are particularly common immedeately overlying the top of the calcisphere wackestones, representing recognised that the 'porcellanous limestones' have limited lateral persistence, and reflected this in his tentative correlations (Somerville, 1979b, fig.1).

2) Somerville recognised 15 cycles in the Tynant Formation as exposed at Tynant, each one defined by an upper phase of Calcisphere Wackestone (MA.1). Detailed analysis of the uppermost cycle and others indicate that some may be considered multicyclic (see figs. 9a & 9b).

3) Somerville noted that the large chonetid Daviesiella llangollensis was 'not known from the overlying Eglwyseg Formation' (Somerville, 1979b, p.397). Chart B attests to its infrequent presence in strata of the first ?and higher Eglwyseg Formation cycles, all along outcrop from Minera to Trefor Rocks in the south.

2.1.1. Cycle Form

In the Tynant Formation the cycles vary between about .50cm and 12m thick. Most are between 1 and 4m (see chart A). Ideally they comprise a basal bioclastic phase of argillaceous alga packstones (MA.2 microfacies association) and an upper phase of calcisphere wackestones (MA.1 microfacies associations) corresponding to the 'porcellanous limestones' and 'calcite mudstones' of previous workers. Intermediate in position within the cycle, thin to medium bedded alga peloid grainstones (MA.3 microfacies association) are variably developed, (figs 11& 12) but develop as thicker units in cycles towards the top of the Tynant Formation, reflecting a gradual transition to cycles of Eglwyseg Formation character.

Microfacies commonly grade into each other. Minor litho-facies including intraformational and extraformational conglomerates occur in basal phases of cycles, and are particularly common immed-eately overlying the top of the calcisphere wackestones, representing

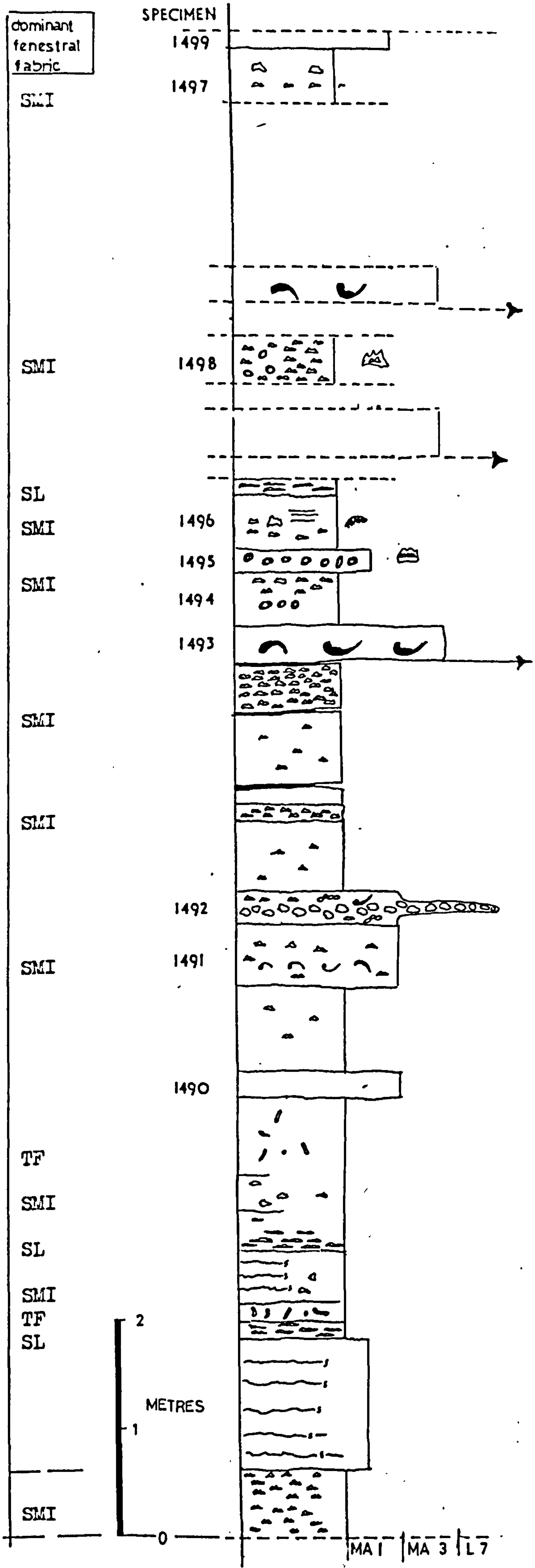
FIG 9a

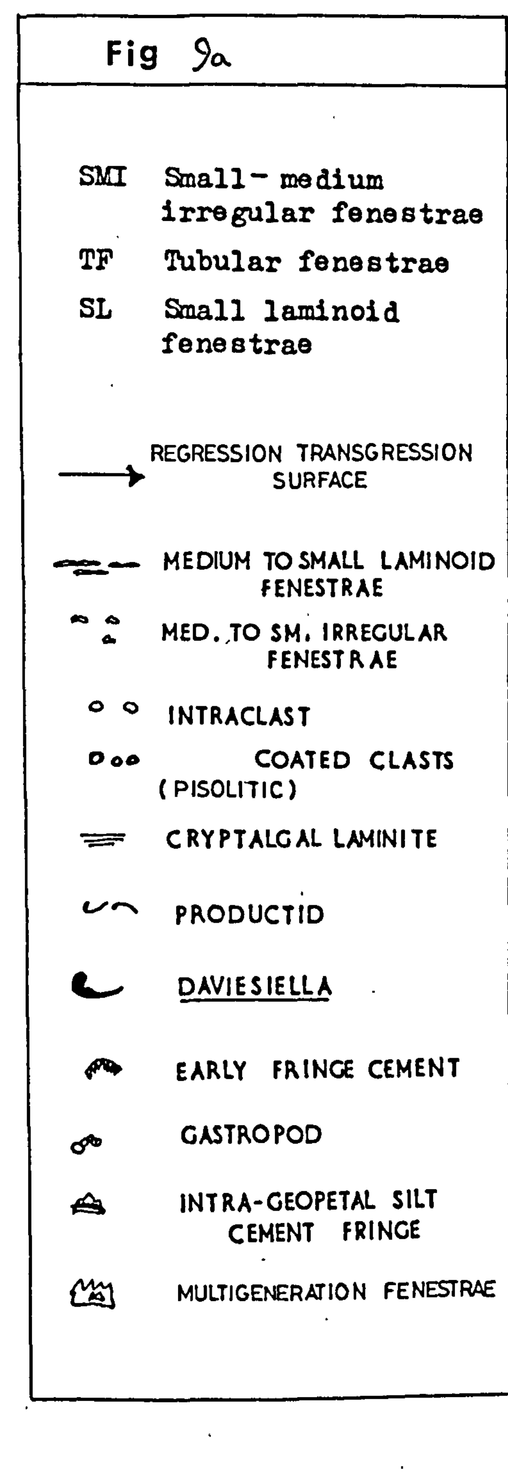
MESOTHEM D5a UPPERMOST REGRESSIVE DEVELOPMENT

-- comprising a sequence of regressive phase peritidal facies dominated minor cycles

TYNANT FORM^{n}

GR SJ 21994561



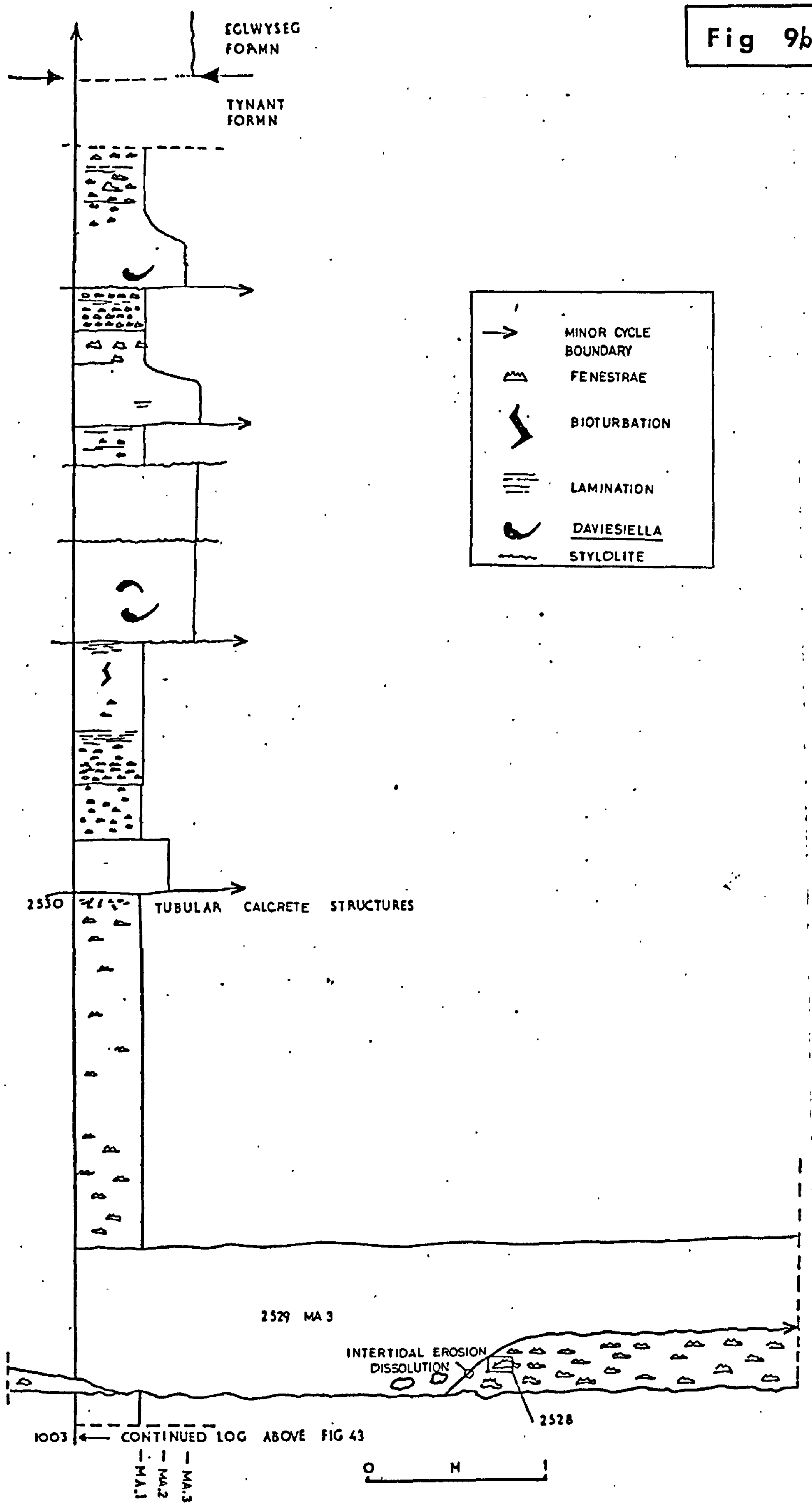


LLWYN HEN PARC

GR SJ 22064587

MULTICYCLIC NATURE OF REGRESSIVE EVENT AT TOP OF TYNANT FORMATION

Fig 9b



the initial transgressive phase of the succeeding cycle. Chapters 6 and 7 discuss the microfacies petrography, early diagenesis and palaeoenvironment implications. Fig.10 shows a composite sequence cycle (sensu Duff and Walton, 1962) from the lower/middle part of the exposed succession at Tynant. Detailed sampling across cycles in the Tynant Formation reveals a number of readily appreciable compositional trends (refer to figs.11 & 12). These are:

1) The relative increase in lime mud towards, and partly within, MA.1 in cycle upper phases.

2) A corresponding increase in calcisphere components into MA.1 compared to the underlying basal phases of the cycles.

3) Calcite cement correspondingly decreases with lime mud increasing.

4) Identifiable peloids and intraclastic micritised fragments decrease slightly in percentage composition from an approximate steady state in MA.2 and MA.3 (about 10-20%) to<10% in upper phases of MA.1 lithologies.

5) Beresellid and tubular / septate algae decrease markedly in abundance into MA.1 lithologies from their domination within underlying MA.2 and MA.3.

6) Foraminifera, red algae and macrofaunal bioclasts are most abundant in MA.2 and MA.3 microfacies, decreasing to trace proportions in overlying MA.1.

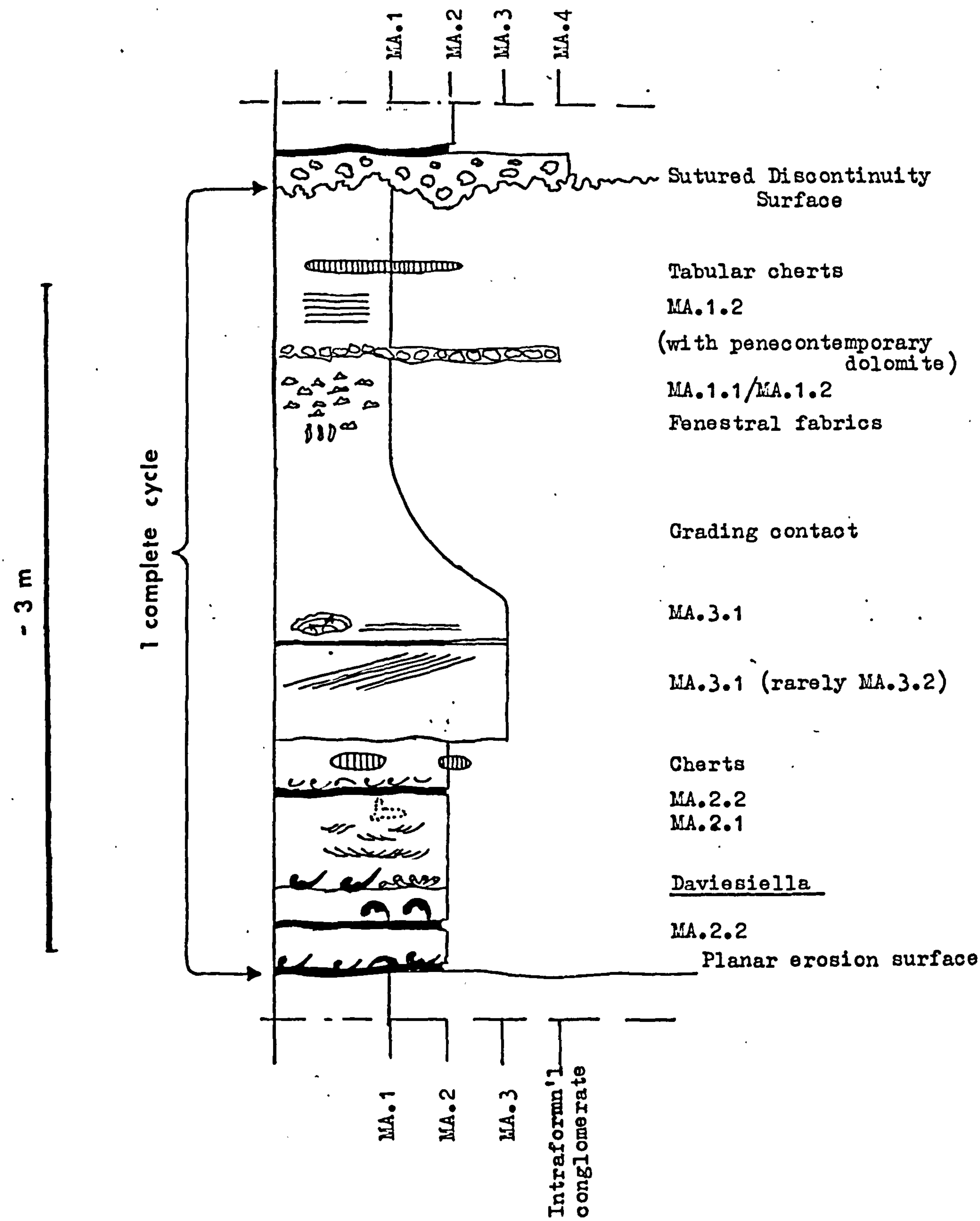
7) Ostracods appear to have an irregular distribution pattern, not bearing on the microfacies type.

These trends are readily recognisable not only in Tynant Formation cycles, but also comparable cycles from the Lower Asbian of the Ravenstonedale area (Potts Beck Limestone, River Cleugh -see fig. 35) section 5. 1. 2 and the Sychtyn Member.

COMPOSITE SEQUENCE TYNANT FORMATION CYCLE

(near Tynant)

Fig 10



Sequential sampling across a cycle and subsequent point counting not only shows these broad trends, but also is a powerful indicator of more subtle 'within-cycle' irregularities and smaller scale cyclic patterns, not readily apparent in the field. Figs. 11 & 12 show where two cycles are differentiated, although only one is readily visible in the field due to the lateral impersistence and poorly developed form of an intermediate calcisphere wackestone microfacies association (MA.1) 'within' the more-readily detected cycle. Other such cases may have escaped detection.

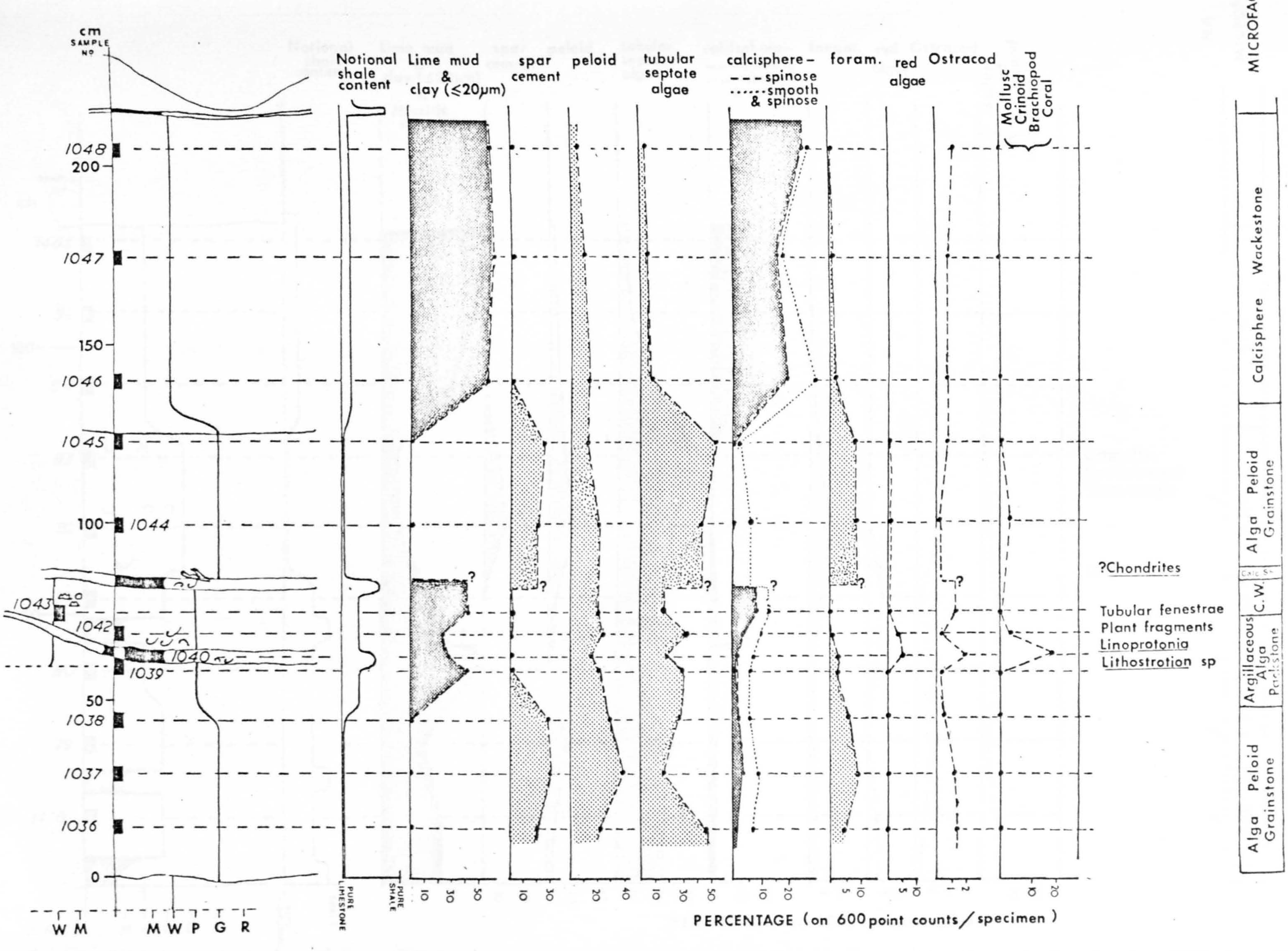
2.1.2. Cycle trend plots ('C-A-M' ternary diagrams)

Fig. 14 shows the compositional variation from point counts of bioclast components in individual samples of Tynant Formation sediments. These ternary plots have been designed from the preceeding review of the main compositional trends within the cycle sediments, in order to 'spread' the sediment plot positions over the diagram as much as possible, whilst allowing for the variability of all the sediments studied. Thus, components with similar trends between microfacies have been grouped as one end member as far as is convenient. This technique, therefore, links components with like depositional histories as single end members.

The three end members erected are:

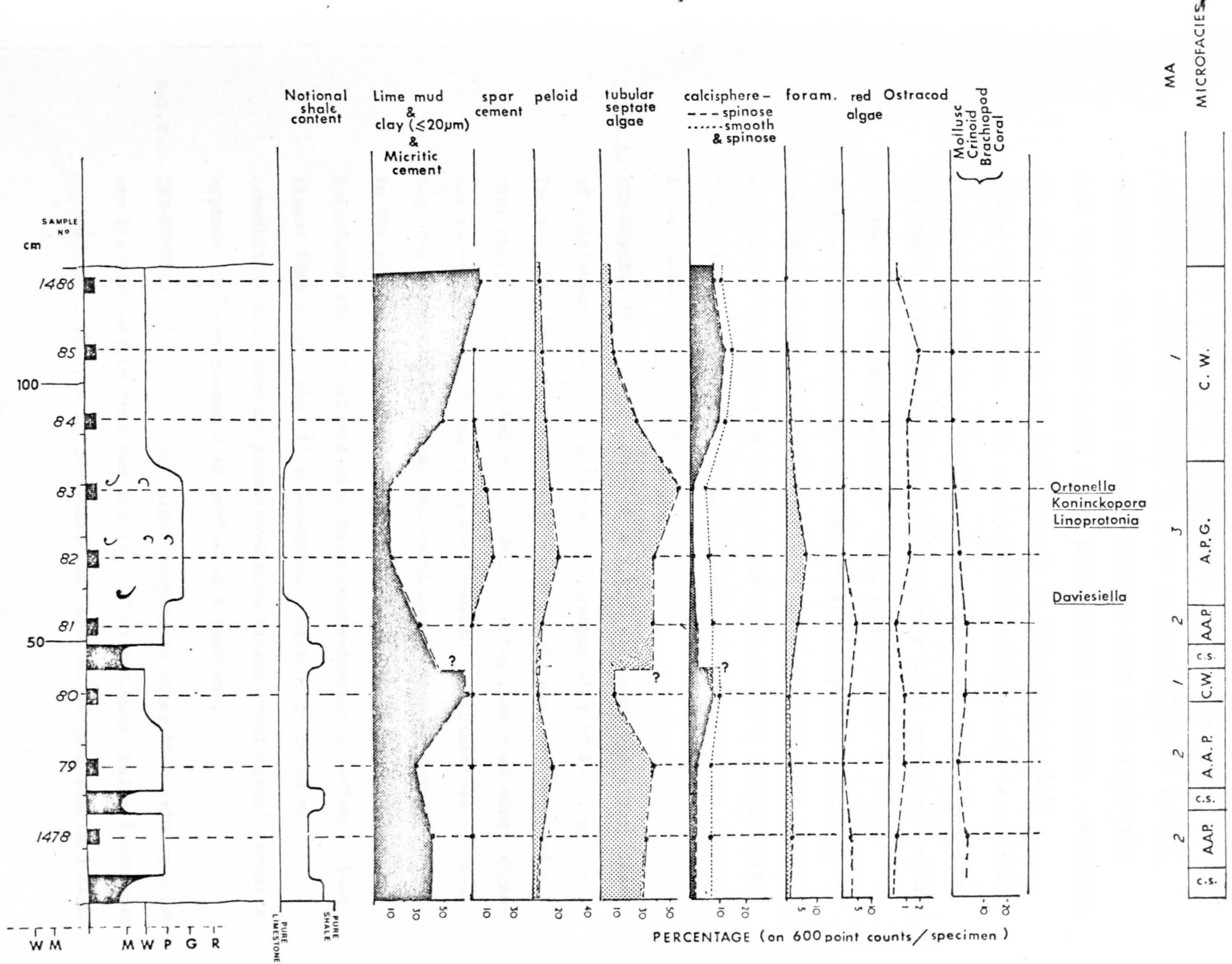
1) Calcispheres + Molluscs (Gastropods) ....(C)
2) All calcareous algae other than calcispheres ....(A)
3) Macrofossil bioclastic elements other than mollusc, but including foraminifera ....(M)

Read (1974, p.27) used tenary plots to distinguish sediment types, but included non-bioclast data. Flugel (1978, p.168) used a circular plot to compare sediment types, but this requires a separate diagram for each sample, although more components may be plotted.



Detailed log of a minor cycle in the Tynant Form'n (g.r. SJ 22002551)

fig 11



Detailed log of a minor cycle in the Tynant Form'n (g.r.22012550)

fig 12

End-member values are obtained by recalculating point-count 'volumes' of the selected components to a percentage of the total end-member component 'volumes' as shown in Appendix II (rows 17-19). Therefore, the total end member components range from a few per-cent to over 80% by volume of the point-counted sediment. Ostracods are not included in the C-A-M diagram as they appear to have a random distribution. The resultingternary composition is a reflect-ion of the available sediment producing organisms and of the sort-ing and transporting processes that may have acted upon the bioclasts as components of muds or sands. As a direct result of this, specific microfacies plot concentrated within certain regions of the ternary diagram but the composition fields do not locate precise microfacies boundaries as the erection of microfacies also involves fabric criteria that do not vary 'in sympathy' with bioclast compos-itions precisely (fig.18).

2.1.2.1. END-MEMBER. 'C' Section 6.6.1.1. discusses the significance of calcisphere dominated sediments (characteristic of MA.1. micro-facies), and indicates that calcisphere plants may have occupied a very shallow subtidal habitat, either in situations that other algae and non-gastropodal fauna could not tolerate, or themselves 'poison-ing' the substrate for other benthonic communities to develop. It is the presence of gastropods to the virtual exclusion of other macrofauna when associated with calcisphere-dominated sediments that places them within the 'C' end-member. Gastropod-dominated communities are commonly associated with Recent 'restricted' (towards hypersaline) environments of peritidal situations.

2.1.2.2. END-MEMBER. 'A' All algal components other than calcispheres are included within this end member. From the microfacies discussion of section 7.2.7.1, it is concluded that beresellid (and similar) algae

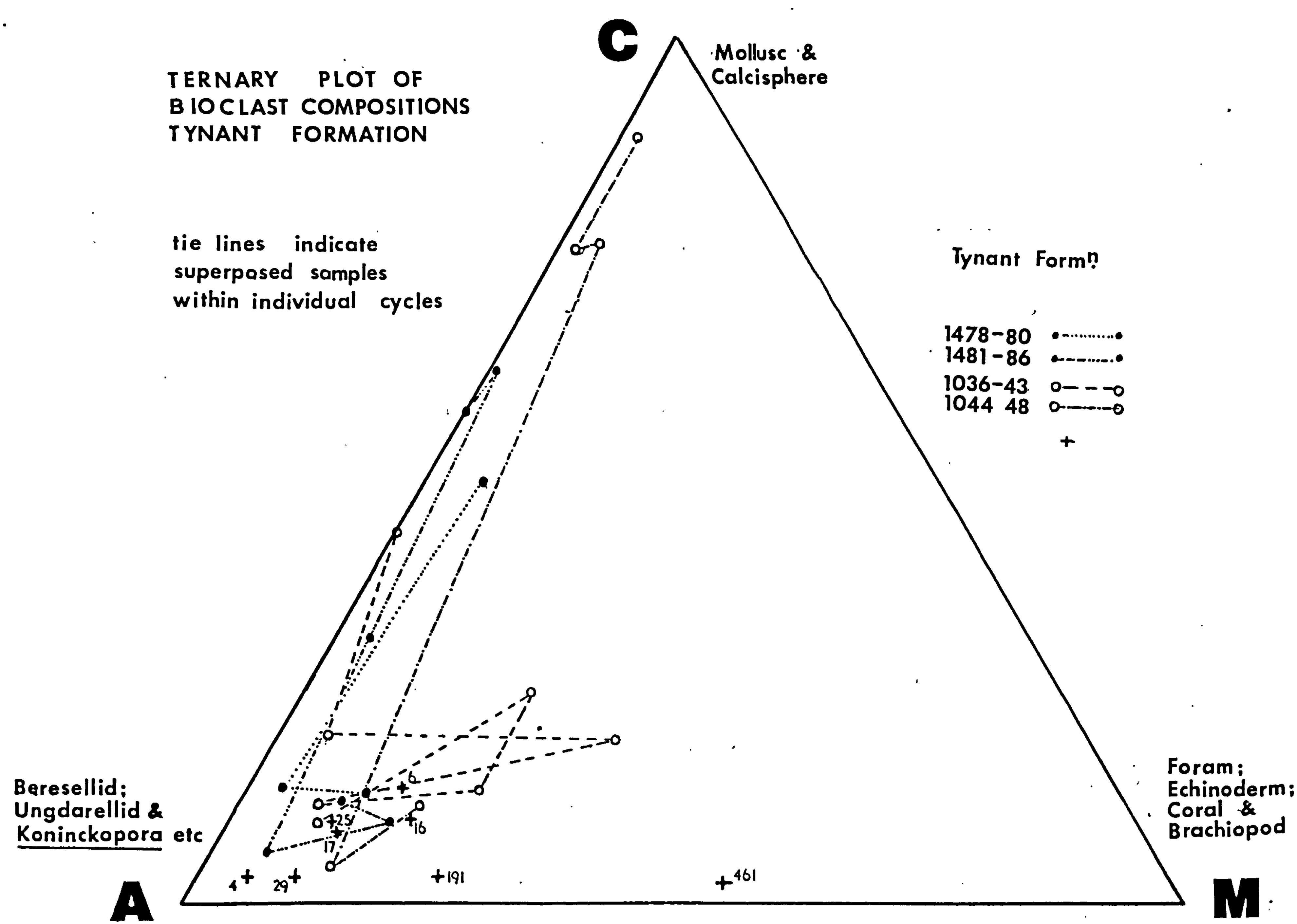

TERNARY PLOT OF
BIOCLAST COMPOSITIONS
TYNANT FORMATION
tie lines indicate
superposed samples
within individual cycles
C
Mollusc &
Calcisphere
Tynant Formn.
1478-80
1481-86
1036-43
1044 48
Beresellid;
Ungdarellid &
Koninckopora etc
Foram;
Echinoderm;
Coral &
Brachiopod
A
M
4
29
191
461
6
25
16
17

FIG 13

were prolific sediment contributors in subtidal environments, below fair-weather wave base where sediment mobility was less than the MA.3 open-shelf habitat. In these sediments the algal diversities are high (see table 5,p110a), and these alga-carpeted habitats supported a variable coral-sponge-brachiopod association.

2.1.2.2. END-MEMBER. 'M' This end-member reflects stenohaline faunal contributions to the sediment (brachiopods; crinoids; corals and bryozoans). Foraminifera are also included here, although their environmental tolerances are not clearly defined. Thus bioclastic grainstones (MA.3.2) plot towards this end-member, as do peloid grainstones in which the recognisable coral+brachiopod+echinoderm + bryozoan + foraminifera bioclast volume far 'outweigh' the algal component.

2.1.2.4. BENEFITS OF THE TECHNIQUE Direct comparisons may be made between the biogenic compositions of sediments, after plotting them on the ternary diagram. Although there is potential for a fourth component it would clutter the diagram, in this case unnecessarily. Wackestones and mudstones plot according to their bioclastic make-up, not their textures. Trends of stratigraphically related samples may readily be picked out, and tie-lines linking them inserted (see fig 13). This technique shows cycle trends, and eliminates total-composition data modified by sediment compaction, matrix and peloid abundance.

2.1.3. Cycle trends in the Tynant Formation.

The ternary diagram , fig .. 13 effectively shows the main Tynant Formation cycle trend, with sediment compositions evolving from a mixed bioclastic composition in basal MA.2 and MA.3 phases towards mollusc-calcisphere dominated MA.1 microfacies on cycle tops. Fig. 17 shows this trend simplified.

Lower cycles of the Tynant Formation have poorly developed MA.3 units, and at times thick sequences of MA.2 sediments. Towards the top of the Tynant Formation, however, the cycles develop thicker middle units of thin bedded MA.3 sediments, a trend that is continued into the overlying Eglwyseg Formation.

The upper cycles of the Tynant Formation are laterally the most extensive, due to successive onlap on to the flanks of the Cyrn y Brain. North of Minera, these upper cycles contain basal phases dominated by bedded MA.3 lithologies with evidence of a greater energy environment indicated by low angle inclined laminated coarse peloidal grainstones lacking indigenous macrofauna (with skeletal oncolites†, extraclasts, and centimetre sized intraclasts) contrasting to the finely bioclastic (mostly < 1mm) typical MA.3 and underlying M.A.2 of the World's End to Trefor Rocks succession. This transition in microfacies form suggests a northward transition from shallow subtidal deposition (at or below fair-weather wave-base; fig 14 &77) (chapter 7) to near shoreface deposits. The regressive MA.1 microfacies are also poorly developed in the upper Tynant Formation around Minera (except the uppermost regressive phase). Shark Bay offers a Recent comparative situation where sublittoral sheets of coarse lithoclastic-skeletal grainstones underlie tidal-flat veneers (Brown & Woods 1974 p328). Purser (1975) interpreted inclined bedding in 'grainy' cycles to be of beach or prograding spit origin. A subtidal situation is favoured here.

2.1.4. Facies Transitions.

Lateral microfacies transitions are readily observed only in MA.1 associations. These transitions take a number of forms:

1) Abrupt lateral transitions due to a cross-cutting bedform, e.g. surface relief of a stomatolitic cryptalgal laminite (see fig.40) or differential erosion on a cycle boundary (fig. 41). These transitions do not imply synchroneity of deposition of the adjacent lithologies.

2) Grading lateral transitions are difficult to recognise in the

† see frontispiece

SUCCESSION BELOW THE UPPERMOST REGRESSIVE PHASE DOMINATED MINOR CYCLE SET
TYNANT FORMATION A COMPARISON OF MINOR CYCLE FORM

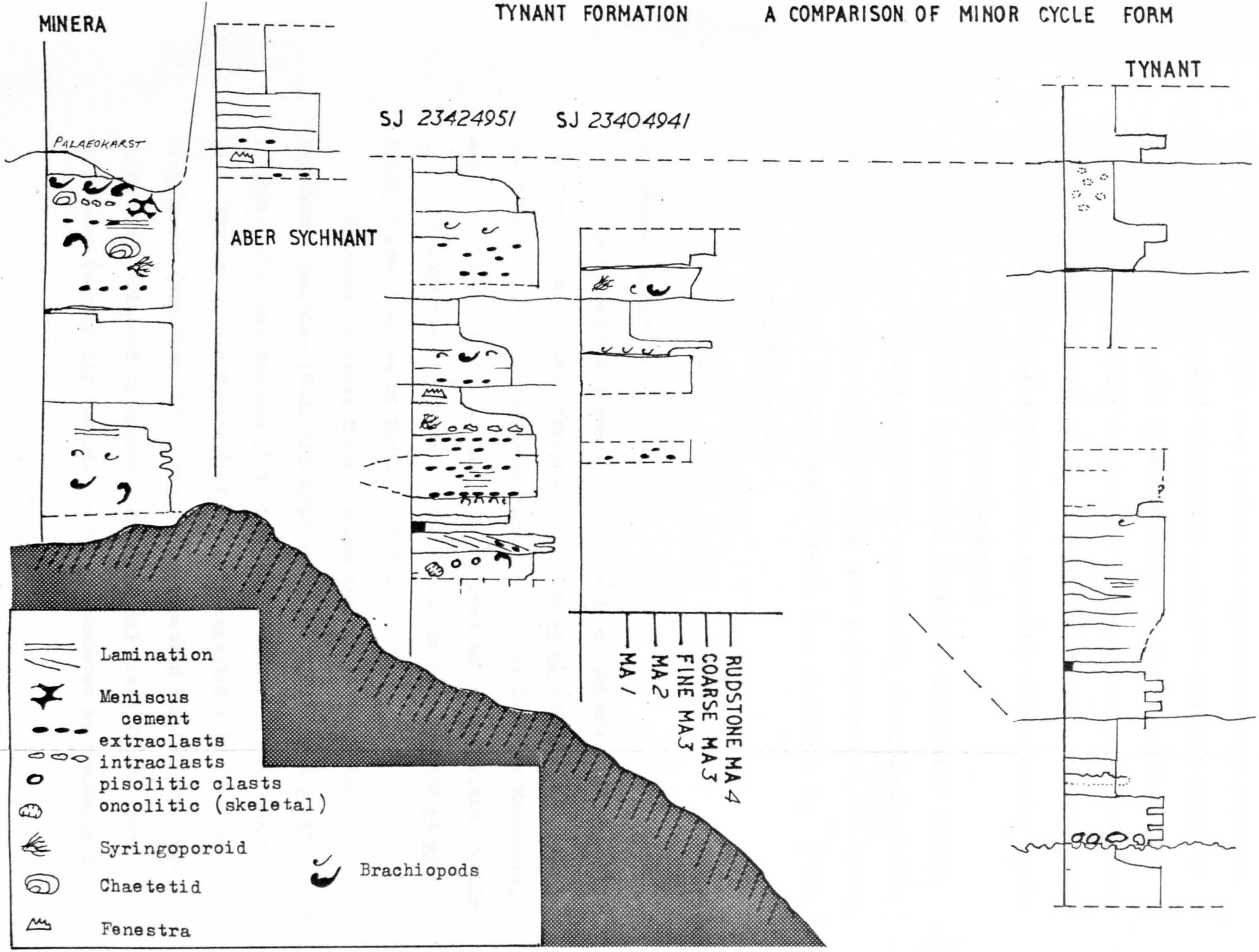


FIG 14

field due to correlation and exposure problems. One rare example of a lateral transition over 5m from an argillaceous alga packstone (MA.2.1) to a fenestrated calcisphere wackestone (MA.1.1) occurs at G.R. SJ. 22002551. The uppermost cycle of the Tynant Formation (according to Somerville's scheme) is characteristic, with a regressive MA.1 phase about 10m thick. However, at Minera, over a distance of 1km, this 10m thick sediment body apparently thins to a few centimetres, but with an overlying palaeokarst.

Lateral variation in internal structures are locally marked. Alga peloid grainstones (MA.3) vary considerably from unlaminated to parallel and to cross-laminated units over a distance of a few metres, due to the presence of localised bedforms (see section 6, 3.) and ?bioturbation homogenisation.

2.2. WHITEHAVEN FORMATION

The WHITEHAVEN Formation (new term) is equivalent to the lower 150 feet of the 'Lower Limestones' of Wedd et al (1929). It also is the lowest carbonate- dominated formation of the Oswestry Embayment, and in a similar manner as the Tynant Formation, is underlain locally by a siliciclastic formation - the 'Basal Shales' of Wedd et al (1929) ('Lower Shales' of Morton , 1879).

Exposure of Basal Shales is now limited to ~2m at the base of Whitehaven Quarries (G.R. SJ264245) - see chart C. Wedd et al (1929) recorded adits dug through 50 feet of Basal Shales at Crickheath Hill (G.R. SJ273233), and Morton (1879, p.121) recorded ~ 62 feet at Llanymynech (see Chart C). Figures approaching 100 feet have been suggested, dominated by maroon, dark grey and green shales, thin carbonates (especially towards the top), arenaceous carbonates, and

MESOTHEM D5a OSWESTRY EMBAYMENT

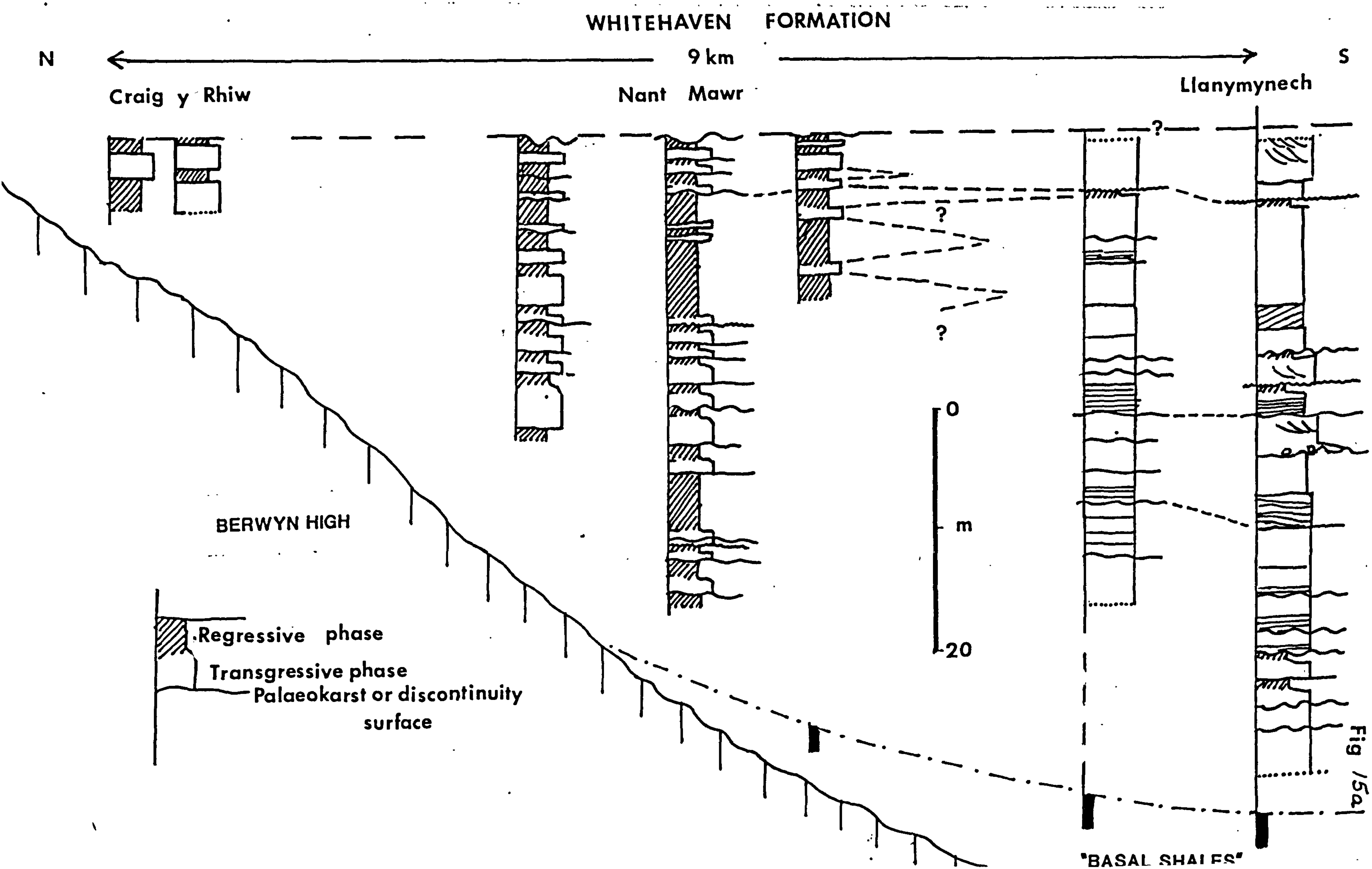


Fig 15a

Lower Palaeozoic extraclastic conglomerates (Morton, 1879, p.120 - 121). Wedd et al (1929) recorded both marine faunas and much fragmentary plant material from the basal shales.

2.2.1. Definition of the Whitehaven Formation. (Fig. 15a)

This formation is composed of the Sychtyn and Llanymynech Members. It comprises a sequence of shoaling cycles with alga-peloid grainstone dominated basal phases and either calcisphere wackestone upper phases or bedded cross laminated grainstones. Cycle boundaries are either emergence surfaces or planar transitions. The formation name is derived from the area of country, north of Llynclys Hill, upon which numerous quarries expose the Sychtyn Member.

The upper boundary of the Whitehaven formation (with the Llynclys Formation) is marked by the commencement of cyclic deposition similar in form to the Eglwyseg Formation, with argillaceous stylonodular and thick bedded MA.3 lithologies predominating. The boundary is a deeply 'potted' palaeokarst (see chart C).

The Whitehaven Formation is divided into the Sychtyn and Llanymynech Members (new terms), each with distinct lithological associations, cycle forms, and outcrop pattern (Fig. 15b).

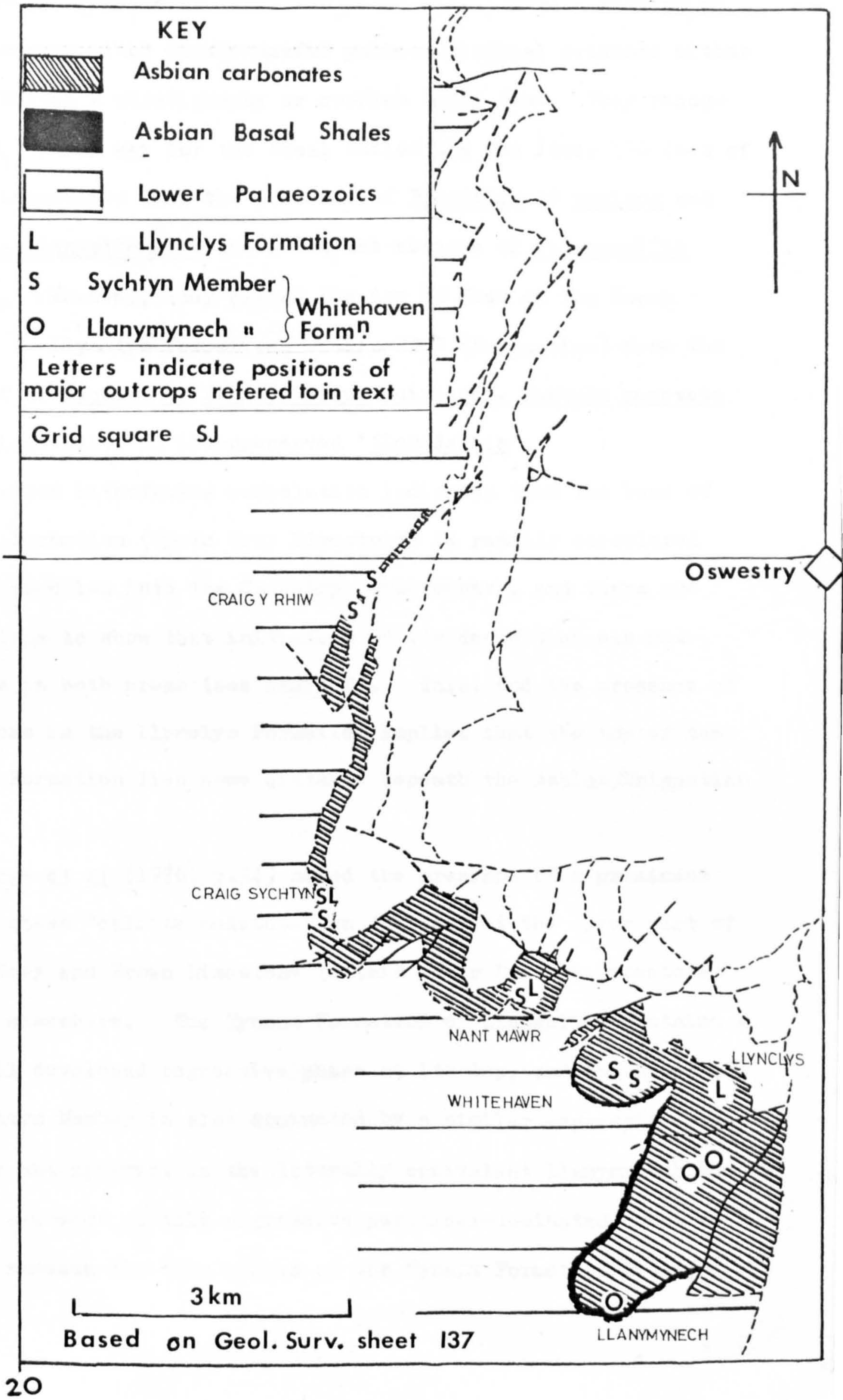
2.2.2. Biostratigraphic controversy.

Morton (1879) correlated (implied chronostratigraphic) the Eglwyseg Formation with the Whitehaven and Llynolys Formations (his "White Limestone") in the Oswestry area. He also placed the "Lower Shale" (op.cit. p.120) as equivalent to the "Lower White Limestone" of the Llangollen area, primarily on account of the relative thicknesses of the Llangollen and Oswestry sediments.

Conversely, Wedd et al (1929, p.89) placed the "Lower Limestones" as equivalent (?chronostratigraphically) to the "White Limestone and greater part of the Lower Grey and Brown Limestone of the Llangollen

Fig 15 b

WHITEHAVEN FORMATION & LYNCLYS FORMATION — OUTCROP & LOCALITIES



area", dismissing Morton's views as "contrary to palaeontological evidence and not justified by lithological resemblance, except in the highest beds [≈Llynclys Formation]" (op cit. p.90). Wedd et al (1929), however, cited little useful palaeontological evidence either to refute Morton's stratigraphy or confirm their own. They recognised a "D_1" zonal age for the Basal Shales and the lower 150 feet of the Lower Limestones with the presence of Productus cf maximus and Daviesiella llangollensis, but noted the absence of Palaeosmilia murchisoni. However, they placed the top 80 feet of the Lower Limestones (≈Llynclys Formation) within "D_2" (Brigantian) from the presence of P. muchisoni, Dibunophyllum muirheadi, Pustula punctata and, most important, an ill-preserved '?Lonsdaleia'.

Detailed lithofacies correlation indicates that the base of the Trefor Formation (Upper Grey Limestone) is readily correlated from the Llangollen into the Oswestry embayments , and there are no indications to show that initiation of its deposition was not synchronous in both areas (see chart D). This, and the presence of Asbian faunas in the Llynclys Formation implies that the top of the Whitehaven Formation lies some distance beneath the Asbian/Brigantian boundary.

George et al (1976, p.34) noted the presence of a prominent regressive phase "calcite mudstone" on Anglesey, at the upper part of the Lower Grey and Brown Limestone (within their Dyserth Limestone Group), and elsewhere. The Tynant Formation of Llangollen contains a similar well developed regressive phase at its top, and the upper 15m of the Sychtyn Member is also dominated by a similar regressive phase, but this is not apparent in the laterally equivalent Llanymynech Member.

The presence of this regressive peritidal-dominated event as a transition between the thin cycles of the Tynant Formation/Sychtyn

Member and the thicker cycles of the Eglwyseg/Llynclys Formation sediments promotes inclusion of the Whitehaven Formation within the early Asbian as lithostratigraphic (and chronostratigraphic) equivalent to the upper cycles of the Tynant Formation (as concluded by Wedd et.al, (1929)). Thus the underlying Basal Shales are chronostratigraphically equivalent to lower cycles of the Tynant Formation.

The "Porcellanous Bed" of the Askrigg Block (section 5, 2.) may be a synchronous early Asbian regressive event (cited in Somerville, 1979b), although Jefferson (1980) considered this and the underlying Horton Limestone as Holkerian, and followed previous workers in evoking a non-sequence spanning most of the early Asbian on the Askrigg Block.

Biostratigraphic correlation of the Whitehaven and Tynant Formations is more tenuous. Daviesiella llangollensis, common in basal phases of Tynant Formation cycles, is a very rare component of the Whitehaven Formation. Gigantoproductus cf maximus, Dibunophyllum sp., Lithostrotion cf decipiens and L. cf. martini occur within the Whitehaven but not the Tynant Formation, and chaetetid-syringoporoid biostromes, a characteristic feature of the Whitehaven Formation are very poorly developed within the Llangollen equivalent. Similar assemblages of algae and foraminifera occur in both formations (see appendix IV), but have not been recognised as distinct from the Eglwyseg Formation. On an assemblage-structure level the Whitehaven Formation has close similarities with the lowermost Eglwyseg Formation. Heterocorals, common in lower cycles of the Eglwyseg Formation, are however, absent in Whitehaven Formation sediments, but occur within the overlying Llynclys Formation. It is here considered that the biostratigraphic differences between the Tynant and Whitehaven Formations are palaeoenvironmentally induced, reflected in the subtle variations of cycle form and palaeogeographic/depositional disposition of the areas (see on).

2.3. SYCHTYN MEMBER (New Term)

Derivation of name: From Craig Sychtyn, the Major outcrop feature of this member.

Type locality: Nant Mawr Quarry, G. R. SJ.252250, near Treflach, 1½ km south of Craig Sychtyn (Chart C).†

Lithostratigraphic definition: The member comprises a sequence of thin shoaling carbonate cycles (<1m to a maximum of about 15m) dominated by a basal alga peloid grainstone phase and an upper calcisphere wackestone phase. It is encompassed within the Whitehaven Formation. Southwards it interdigitates with the Llanymynech Member whilst northwards it onlaps successive cycles on to the Berwyn 'axis'. (see Chart C).

Thickness and distribution. It varies between zero over the Berwyn Hills to a maximum of about 50m in the Whitehaven area.

The Sychtyn Member outcrops between Craig y Rhiw in the North and Whitehaven in the South. It is onlapped towards the Berwyns by the Llynclys Formation (in turn

† Plate 1, fig. E.

PLATE 1. Mesothem. D5a

A Lensing horizon of calcisphere wackestone (MA.1) =C.W., within the Tynant Formation, Tynant, G. R. SJ 21994551. Note 'sutured discontinuity' upper surface.

B A 15cm bed of matrix supported intraformational conglomerate (C) overlying a minor cycle boundary (B outlined) at Tynant, Tynant Formation, G. R. SJ 22054533.

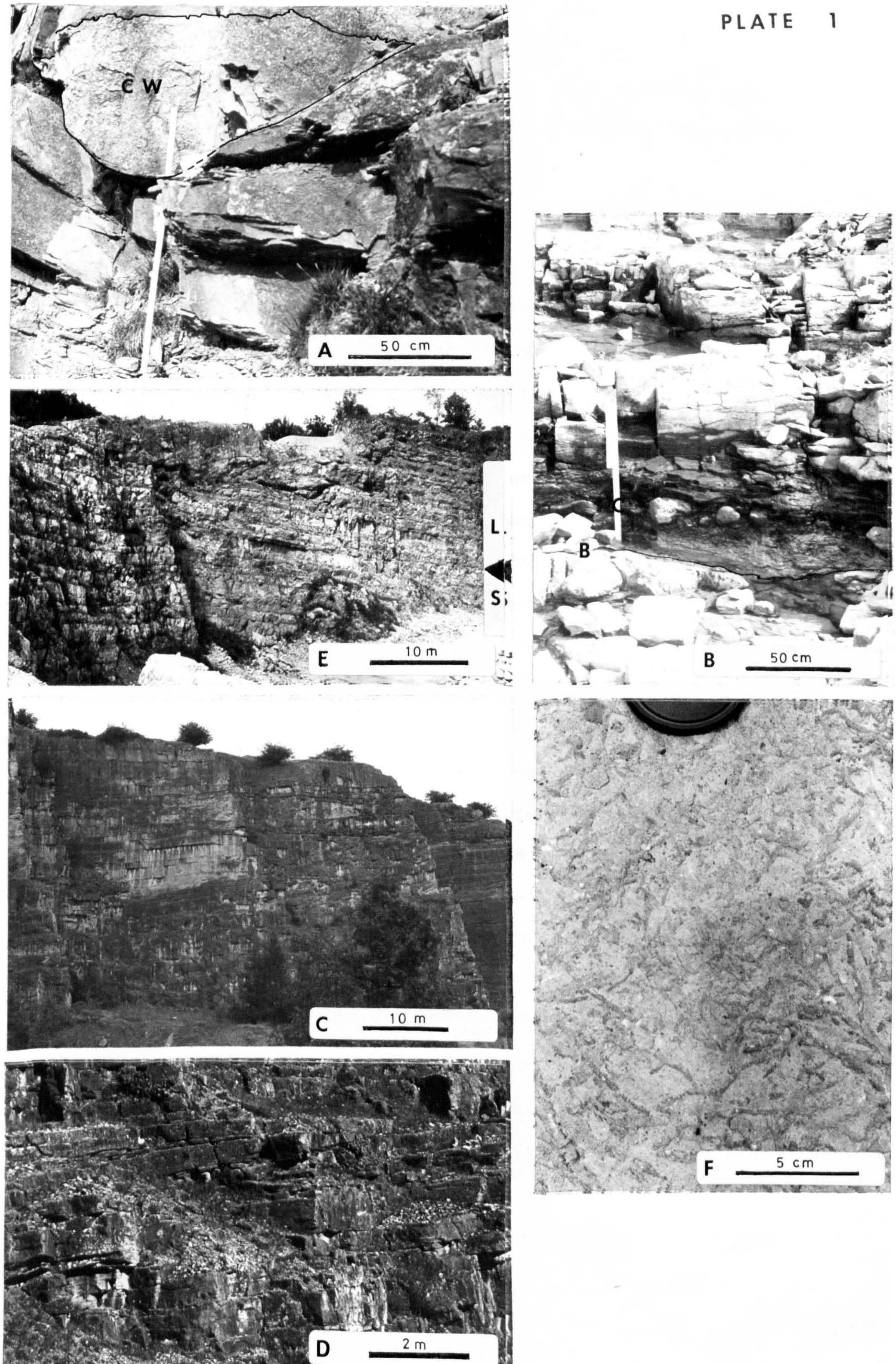
C Quarried exposure of Asterley Rocks, Llanymynech Hill, Whitehaven Formation, Llanymynech Member, G. R. SJ 267218.

D Llanymynech Member, Llanymynech Hill, "Old Quarry", G. R. SJ 26572180. Bed lensing within a low angle cross bedded unit.

E Type section of Sychtyn Member (Whitehaven Formation) and lower minor cycles of the Llynclys Formation (upper Asbian) Nant Mawr Quarry, G. R. SJ 25202500.

F Chondrites sp. within an alga peloid grainstone (MA.3) Sychtyn Member, Nant Mawr Quarry, G. R. SJ 25122503.

PLATE 1



onlapped by the Trefor Formation). It is underlain by the Basal Shales only south of ?Treflach, north of which it oversteps the Lower Palaeozoics of the Berwyns.

2.3.1. Lithostratigraphy (refer to Chart C.)

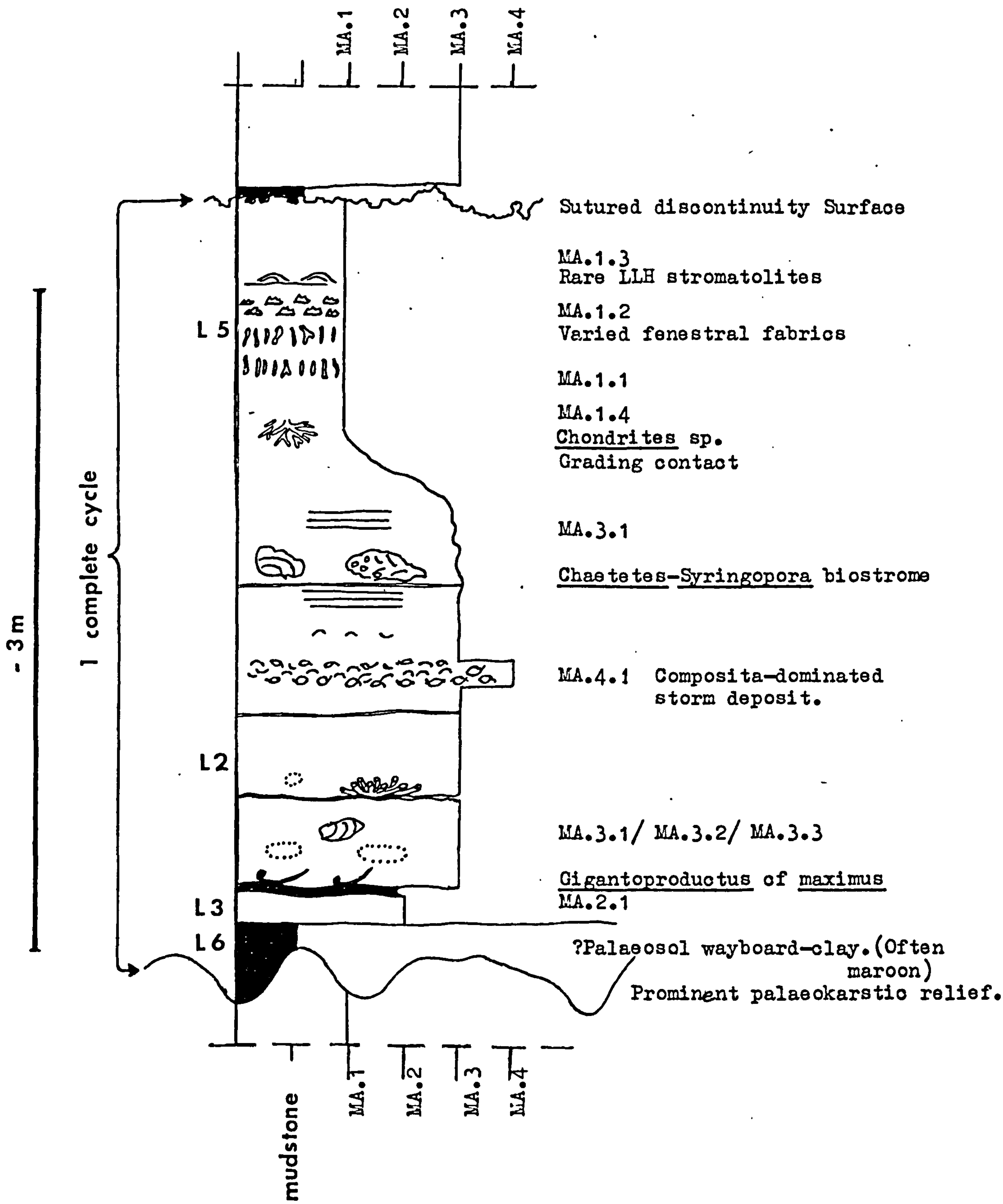
The Sychtyn Member has a cyclicity of form similar to the Tynant Formation. However, algal packstones, (MA.2) the dominant microfacies association of the Tynant Formation, are subordinate to peloid-algal grainstones (MA.3), forming most of the cycle lower phases. Calcisphere wackestones (MA.1) account for up to 50% of the formation. The cycles vary from 30cm to<15m in thickness† and the sediments are generally less argillaceous, and lighter grey/cream or maroon colour compared to the Tynant Formation cycles. A minimum of 15 cycles is readily recognised at Nant Mawr, but the total may exceed 25, as the recognition of cycle boundaries is uncertain. Some may be interpreted as multiple; on others paleokarstic/erosional features are not apparent. As with the Tynant Formation, lateral correlation is effective only in very adjacent outcrops (e.g. within the Whitehaven Quarry complex), with some single cycles correlateable over > 1km, but only on gross morphological detail (e.g. a thick MA.1 upper phase, capped by a prominant palaeokarst.) Indeed, here, as in the Eglwyseg Formation, the character of the cycle boundaries provides useful correlation markers, although these do vary from deep potted surfaces to smooth undulose bedforms with negligible relief (see chapter 11) even along individual planes.
Wedd et al (1929) recognised one prominent "potted" horizon in Whitehaven Quarry. Notably, calcrete profiles have not been observed in Sychtyn Member sediments.

Towards the top of the Sychtyn Member, a thick calcisphere wackestone (MA.1) phase is developed, similar to the top of the Tynant Formation, but is nevertheless composed of ?2 to 6 cycles (FIG/6b, CHART C).

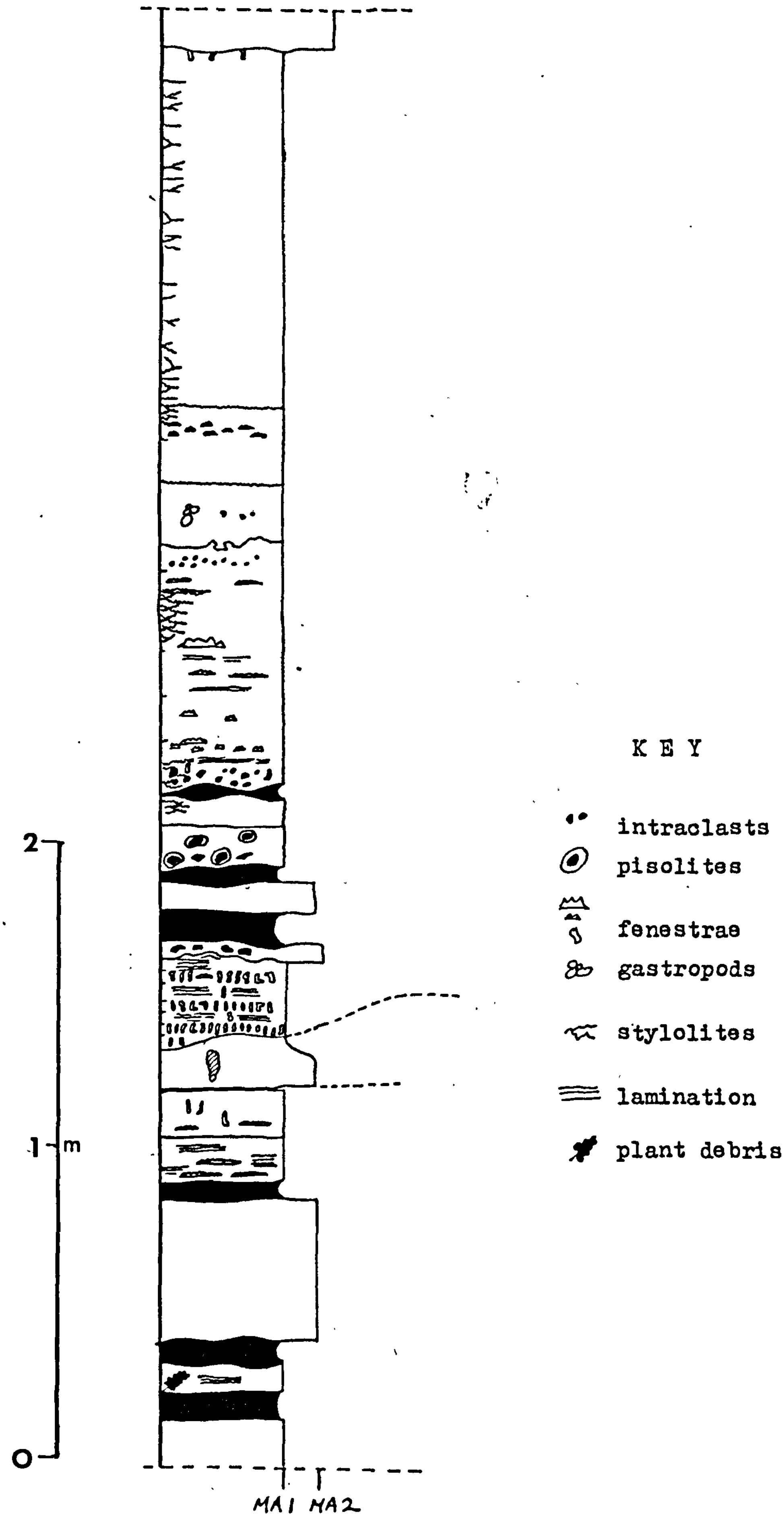
† see composite sequence cycle of Fig. 16a

COMPOSITE SEQUENCE, SYCHTYN MEMBER CYCLE

Fig 16a



Log of part of uppermost regressive-dominated cyclic sequence, Sychtyn Member, Nant Mawr.



The development of MA.3 phases, rarely with stylolitic rubbly weathering and mottled 'pseudobreccia' appearance shows similarities with the Eglwyseg Formation, although these are more apparent in the overlying Llynclys Formation.

2.3.2. Cycle Trends.

Point count analyses of Sychtyn Member sediments reveal very close similarities with similar microfacies of the Tynant Formation (see Appendix II). Their bioclast compositions fall into the same cycle trends as the Tynant Formation (fig. 16c).

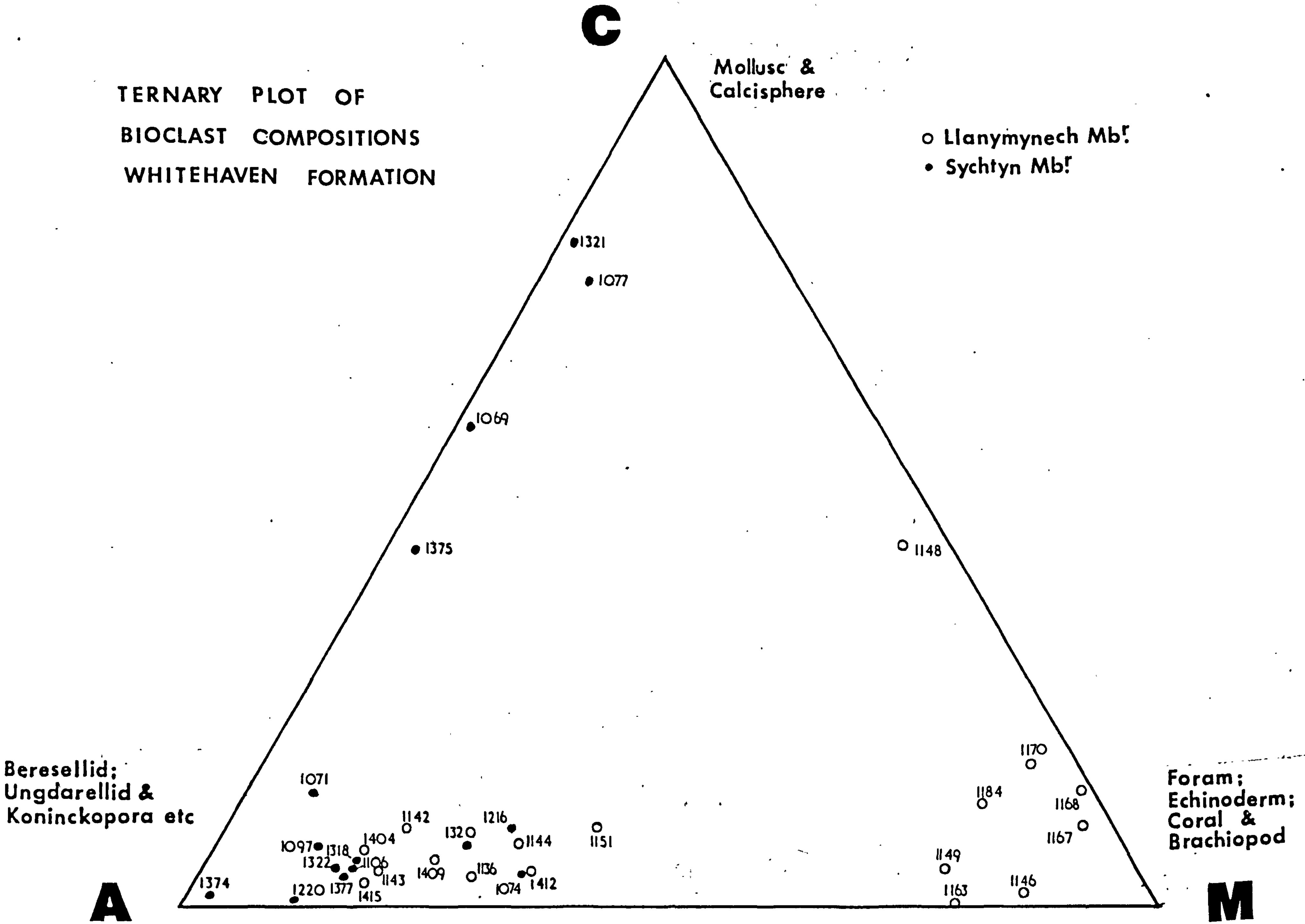
2.3.3. Palaeontology.

Appendix IV lists the fossils associated with the Sychtyn Member. Rugosan corals are rare, and include Lithostrotion cf martini. The presence of Gigantoproductus cf maximus, Daviesiella llangollensis and Linoprotonia hemisphaerica confirm an early to mid Asbian age. However, Composita cf ficoides, an Holkerian form according to George et al. (1976, p. 9 - 11) is very common.

2.3.4. Palaeoenvironment.

The palaeoenvironment interpretations of Sychtyn Member microfacies are discussed at length in chapters 6 and 7. The cycles represent regression and tidal flat progradation, following rapid transgressive events, during a gradual transgression on to the flanks of the Berwyn 'axis', associated with the lower Asbian (Mesothem D5a of Ramsbottom, 1977) transgressive event.

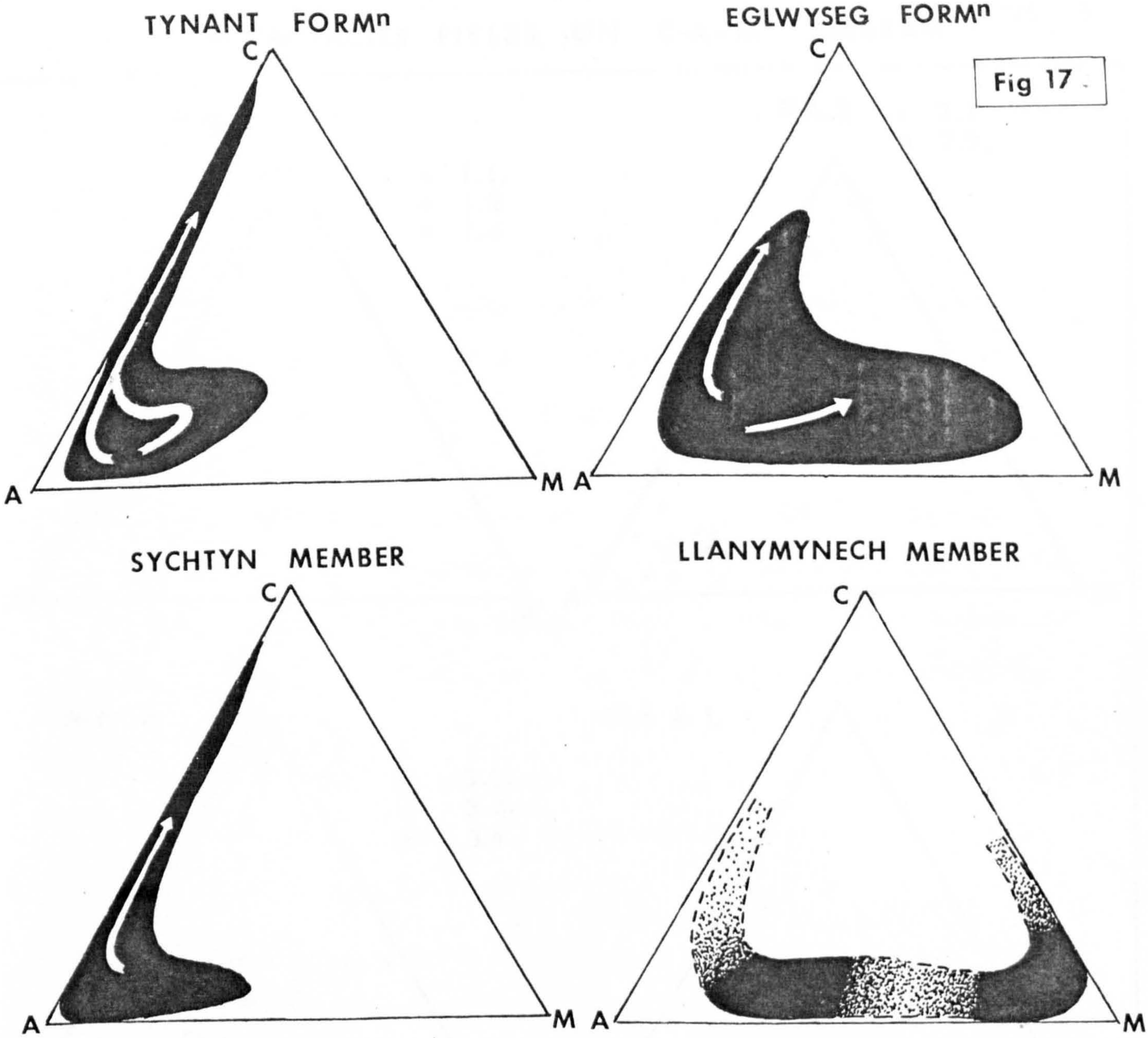
The lack of MA.2 transgressive phases suggests that during maximum transgression, deposition continued above a normal (fair weather) wave base.



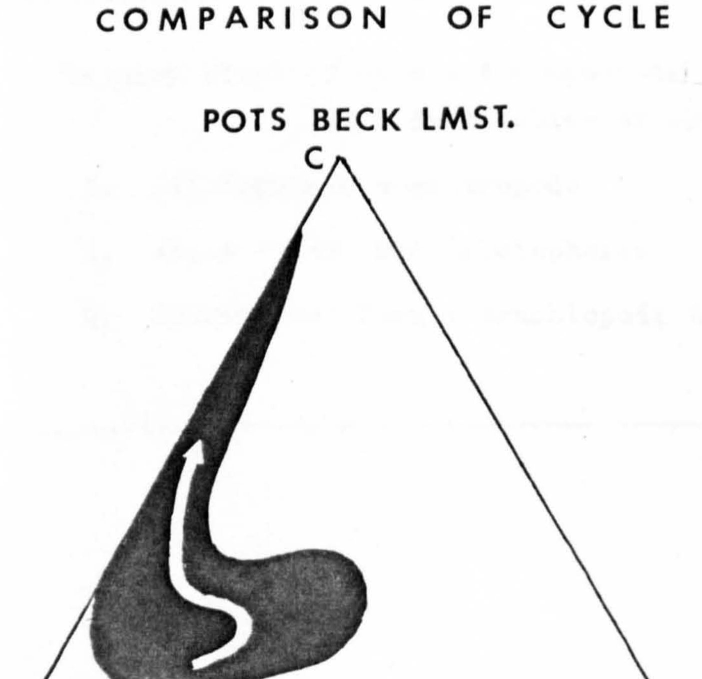
TERNARY PLOT OF BIOCLAST COMPOSITIONS WHITEHAVEN FORMATION

FIG 16c

Fig 17



COMPARISON OF CYCLE TRENDS (ASBIAN)



Ternary plots of bioclast composition within Asbian sediments. Note wide spread of compositions within the Eglwyseg Formation.

C. Calcisphere + gastropod

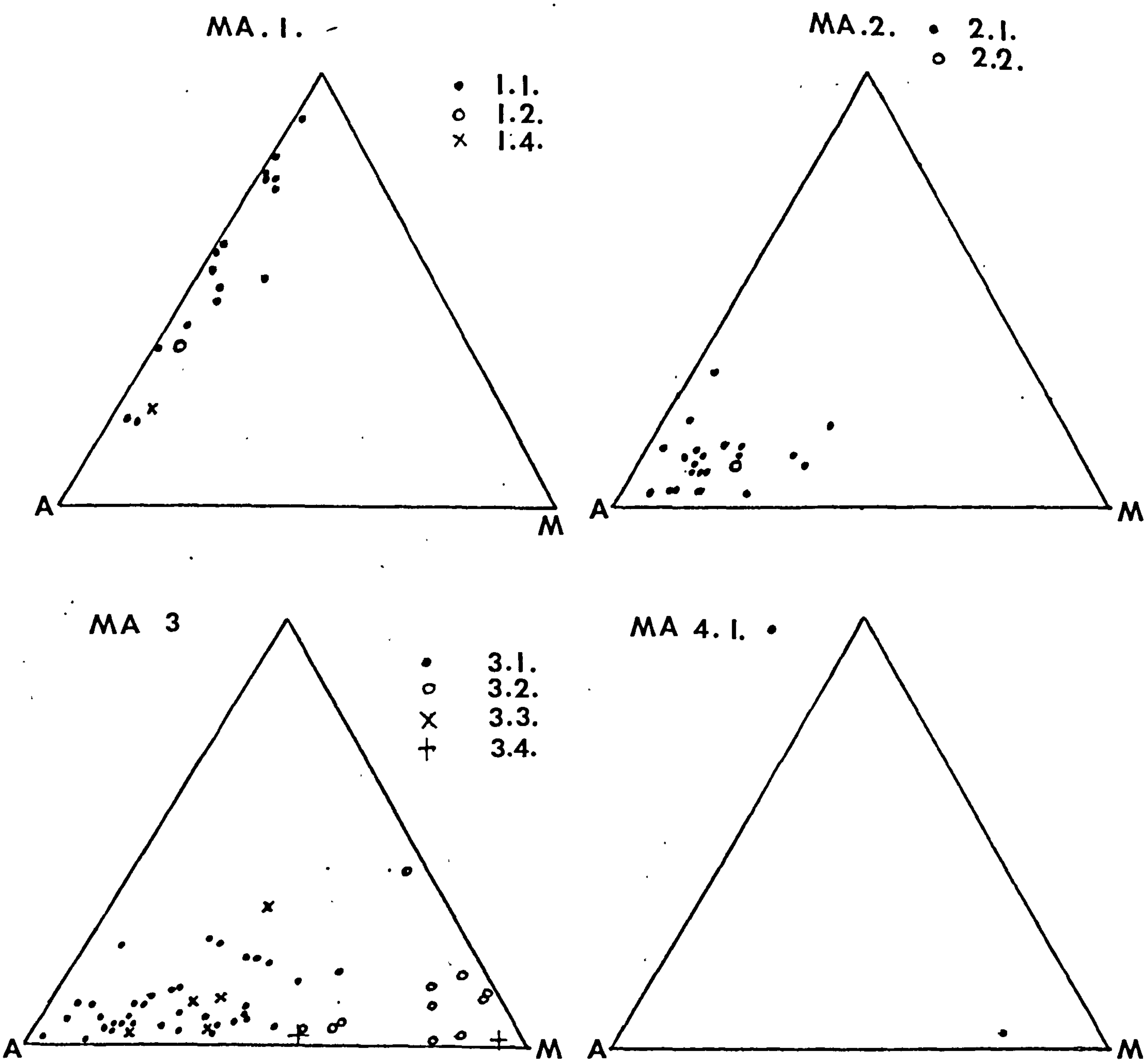
A. Other algae

M, Echinoderm + coral + brachiopod + bryozoan + foraminifera.

Arrows signify dominant trends recognised within individual minor cycles.

MICROFACIES FIELDS ON C-A-M DIAGRAM

FIG 18



Ternary Plots of bioclast compositions within Microfacies Associations.
(Data from tables of point counts, Appendix II.)

C. Calcispheres + gastropods

A. Algae other than Calcispheres

M. Echinoderm; Coral; Brachiopod; Bryozoan; Foraminifera.

2.4. LLANYMYNECH MEMBER (New term)

Derivation of name: From Llanymynech Hill, the major outcrop feature of the Llanymynech Member.

Type locality: Asterley Rocks, G.R. SJ 26572180, (where it is best exposed in the "Old Quarry") on the southern flanks of Llanymynech Hill. (Plate 1, fig. C; Fig. 15b).

Lithostratigraphic definition: The Llanymynech Member comprises a varied sequence 55 to 60m thick, of shoaling carbonate cycles, underlain by the Basal Shales (unexposed) and overlain by the Llynclys Formation (transition unexposed). It is encompassed within the Whitehaven Formation and?interdigitates with the more northerly Sychtyn Member. The minor cyclicity lacks well developed regressive calcisphere wackestone (MA.1) phases, and is dominated by a varied sequence of shoaling grainstones interspersed with palaeokarsts.

2.4.1. Lithostratigraphy (Refer to Chart C.)

The Llanymynech Member outcrops on Crickheath and Llanymynech Hills. It is chronostratigraphically equivalent to and apparently interdigitates with the Sychtyn Member. It comprises~90% MA.3 (alga peloid grainstone) microfacies, mostly in a cycle form distinctly different to the Sychtyn Member, with only very rare and poorly developed MA.1 upper, and MA.2 lower phases. About 15-20?cycles are exposed at the type locality, although their individual recognition is often ambiguous. Chart C shows the tentative correlations between the Sychtyn and Llanymynech Members.

Whereas cycles of the Sychtyn Member characteristically had MA.1 dominated upper phases, often with emergence features developed thereon, cyclicity in the Llanymynech Member is less readily recognised. It comprises basal MA.3 (and MA.2) phases of thin bedded finely bioclastic-peloidal sediments, at times with large scale low angle (2-15°) cross-bedding (Plate 1 fig. D), and 'pinch and swell' structures, superseded by MA.3 sediments in beds up to 3m thick, often with oolites and internal cross-lamination common(fig. 57).

Cycle boundaries are poorly defined, either by marked lithological transitions, or by palaeokarstic and ?palaeokarstic surfaces (Forms 'A' and 'D' - see section 11.2.1.). Calcrete phenomena are rare. Often only an 'upper' massive bedded MA.3 phase is present sandwiched between (?) emergence surfaces and, therefore, representing an individual cycle. Chapter 8 discusses the microfacies relationships and palaeoenvironment model for these deposits. Fig. 19 summarises the cycle form.

2.4.2. Cycle Trends.

In a similar manner to Sychtyn Member and Tynant Formation sediments their bioclastic compositions plot on the C-A-M ternary diagram (figs.17 &18). In this Member, the cycle form shows a distinctly different trend. Although some of the thin-bedded basal phases plot towards the 'A' end-member, upper phases dominantly trend towards the 'M' rather than 'C' end member.

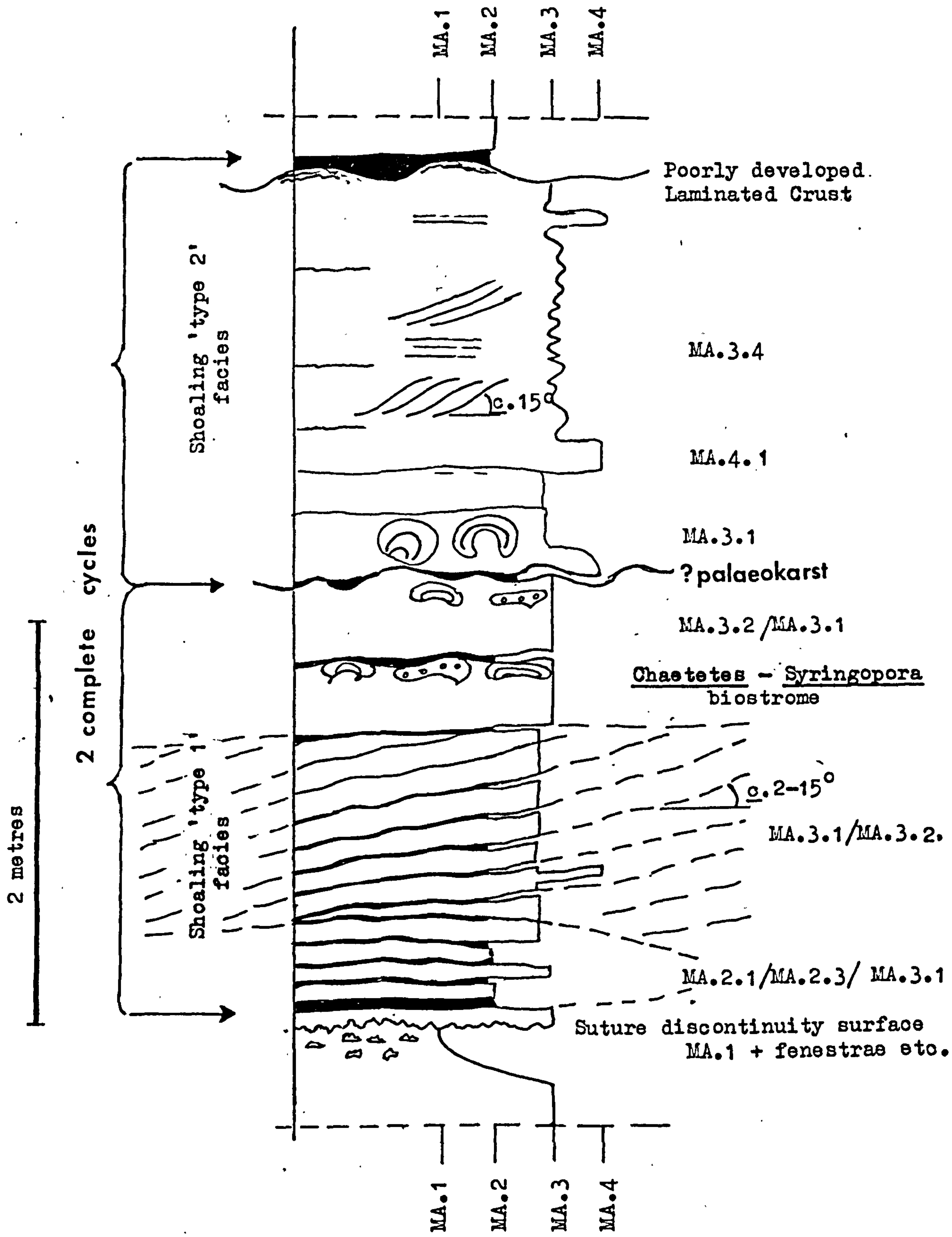
An approach to the 'C' end member is made by the very few MA.1 phases present on cycle tops.

2.4.3. Palaeontology (see fossil list, appendix IV)

2.4.3.1. MACROFAUNA Palaeontologically, this member is similar to the Sychtyn Member with characteristic Chaetetes(Boswellia)/ Syringopora reticulata biostromes, and similar Composita-prolific brachiopod assemblages. Wedd et al (1929) reported the presence of Daviesiella llangollensis at Llanymynech, which has been confirmed in this study. Lithostrotion cf decipiens occurs in higher beds of the

COMPOSITE SEQUENCE LLANYMYNECH MEMBER CYCLES

Fig /9



Llanymynech Member, and is common in mid/lower cycles of the Eglwyseg Formation of Llangollen.

Dibunophylloids, caninids and fasciculate litho--strotiontids further suggest an Asbian age (see fossil list, appendix IV). Both gastropods and echinoid spines are locally significant.

2.4.3.2. MICROFLORA Microfloral assemblages similar to the Sychtyn Member (Appendix IV) occur, with Ungdarella sp and Asphaltina sp (s/m 1146)

2.4.3.3. MICROFAUNA Foraminifera are ubiquitous to all MA.2 and MA.3 sediments, and include: Pseudoendothyra sp. (1409); rare archaediscids; Plectogyra spp., ?Palaeotextularia sp. (1241); Endothyranopsis cf crassa; Ammodiscids indet. (1415); Eostafella cf proikensis (s/m 1163).

2.4.4. Palaeoenvironment.

Llanymynech Member deposits are the southernmost remaining Asbian sediments of the Oswestry Embayment. To what extent the Carboniferous sedimentation extended southwards (on to St. Georges Land) is unclear but it was probably kilometres rather than 10's of kilometres.

There is, however, no evidence of southerly thinning of the Whitehaven Formation (cf. the Brigantian Trefor Formation).

The sedimentology and palaeoenvironment interpretation are given in Chapter 8. The Llanymynech Formation represents a subtidal shoaling complex, located seawards of the Sychtyn Member, to which it acted as a barrier, especially during regressive phases of the minor cycles, when southerly prograding Sychtyn Member peritidal (MA.1) sediments developed shorewards of it. The overall dimensions and geometry of the Llanymynech Member are not clear. Present outcrop is limited to an elongate NNE/SSW feature 2km by 1km. The development of a 'stacked' shoaling complex may have been caused by an inertial influence of slight topographic relief formed due to the presence of underlying shoaling sediments. Lower minor cycles lack the shoaling developments, and have similarities with the Sychtyn Member sediments, indicating a probable northward progradation of the member reflecting the gradual transgression over the Berwyn high.

2.5. COMPARISONS BETWEEN TYNANT AND WHITEHAVEN FORMATIONS

2.5.1. Lithological comparisons.

The Tynant Formation sediments are mostly darker and more argillaceous than their Whitehaven Formation counterparts. This is reflected in their respective cycle forms. Basal minor cycle argillaceous algal packstone (MA.2) phases dominate the Tynant Formation (at Tynant) whilst bedded grainstones are the Sychtyn Member equivalents. As such, the Sychtyn Member cycles are more similar to the upper cycles of the Tynant Formation developed towards Minera over the Cyrn y Brain (Chart A).

Whereas 'maroon' argillites, from erosion of a Lower Palaeozoic hinterland, provided a clastic supply for the Whitehaven Formation, illitic grey argillites pervaded the Llangollen embayment.

Subsidence probably played a major role in both development of cycle form and overall thickness of the succession. Chapter 3 shows that the basal thin bedded phases of upper Asbian Eglwyseg Formation cycles thicken towards depocentres and are probably related to positions of greatest relative subsidence. It is unclear how much the Llangollen embayment subsided during deposition of the Tynant Formation, as pre-Asbian landscape relief may have partly accounted for the great discrepancy in thickness between Minera, Tynant and the Berwyns (up to about 110m).

Local subsidence apparently played a more subdued role in the Oswestry Embayment. In the 20m below the top thick regressive phase at Nant Mawr (itself 2 - 6 cycles), the Whitehaven Formation comprises about 12 minor cycles. The 12 minor cycles below the equivalent regressive phase of the Tynant Formation account for about 40m of the measured section, suggesting that overall subsidence and/or deposition rates were twice that of the Whitehaven Formation. Within these 12 cycle sets, 730cm of MA.1 phase is developed in the Llangollen embayment (at Tynant) and 860cm in the Oswestry embayment (at Nant Mawr). It is primarily the lack of the basal argillaceous and packstone phases that produce the thickness difference between Tynant and Nant Mawr. Both these locations correspond approximately to the embayment 'depocentres' as exposed at present.

The lack of extraformational conglomerates within the Whitehaven Formation (of the Tynant Formation at Minera) suggests that the Lower Palaeozoic hinterland was either not as proximal, or more subdued and subject to less erosive activity. The former is most likely as extraformational conglomerates have been recorded in the underlying Basal Shales.

The Whitehaven Formation has well developed palaeokarsts on minor cycle boundaries, a rare occurrence in the Tynant Formation. This suggests that longer periods of subaerial emergence characterised the Oswestry embayment, reflecting the absence of local subsidence. Significantly calcretes are rare in both formations.

2.5.2. Biostratigraphic Comparisons.

There are major biostratigraphic differences in the faunal development of Tynant and Whitehaven Formation strata. This difference first prompted biostratigraphic correlation of the Whitehaven Formation with the lower cycles of the Eglwyseg Formation (as Morton, 1879).

2.5.2.1. PALAEOECOLOGICAL SIGNIFICANCE OF DAVIESIELLA DISTRIBUTION.

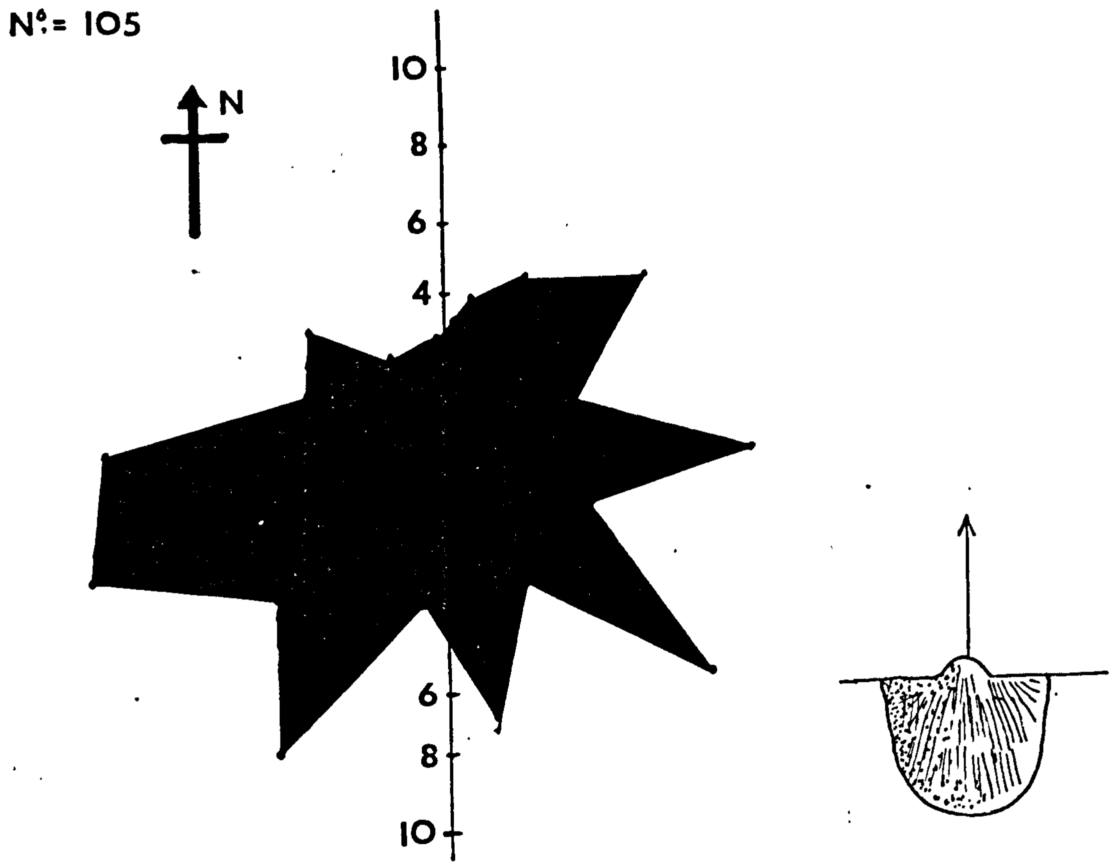
Daviesiella llangollensis, is prolific in the Tynant Formation but a rare component of the Whitehaven Formation. Chaetetid-syringoporoid-lithostrotiontid assemblages are common in the latter, but are scarce in the former.

Substrate sediments in the Oswestry area were more mobile towards cycle bases than Llangollen, possibly reflecting a shallower environment at maximum minor cycle transgression. The formation of oolitic shoals attests to this movement. From the growth morphologies of the chaetetids (section 8.3.1.1) it would appear that they could thrive in a high energy scouring environment. Daviesiellids are large chonetids lacking prominent anchorage spines. Although their robustness and size may have prevented minor bottom currents from overturning them, they (and especially their spat) would have been readily moved above a fair weather wave base. Their adaptive strategy, therefore, was more conducive to a habitat beneath fair-weather wave base, where they had a degree of stability against short lived and minor bottom currents. Fig. 20 shows a rose orientation on directions of umbones of 100 Daviesiella valves from an exhumed bedding plane at Minera. The measurements indicate a degree of orientation towards the SW. (The readings included both convex and concave up single and articulated valves).

Ferguson (1978) showed that gigantoproductids (similar to daviesiellids in gross external morphology) were reorientated by slow moving bottom currents, but were overturned with currents above 30cm/sec. According to Hardie and Ginsburg (1977) bottom current velocities of 100cm/sec are induced by wind speeds of 15m/sec in water 5m deep with 100km fetch. i.e. Storm conditions were not necessary to overturn a daviesiellid in shallow waters.

Further factors controlling the distribution of daviesiellids may have included a degree of environmental restriction (and metahaline conditions) in the Oswestry area, although the coral faunas do not support this, and there is no evidence of any evaporite suite in either the Llangollen or Oswestry embayments.

FIG 20



DAVIESIELLA LLANGOLLENSIS UMBONE
ORIENTATIONS TYNANT FORM'N
MINERA

3. STRATIGRAPHY AND CYCLE FORM OF THE EGLWYSEG AND LLYNCLYS FORMATIONS

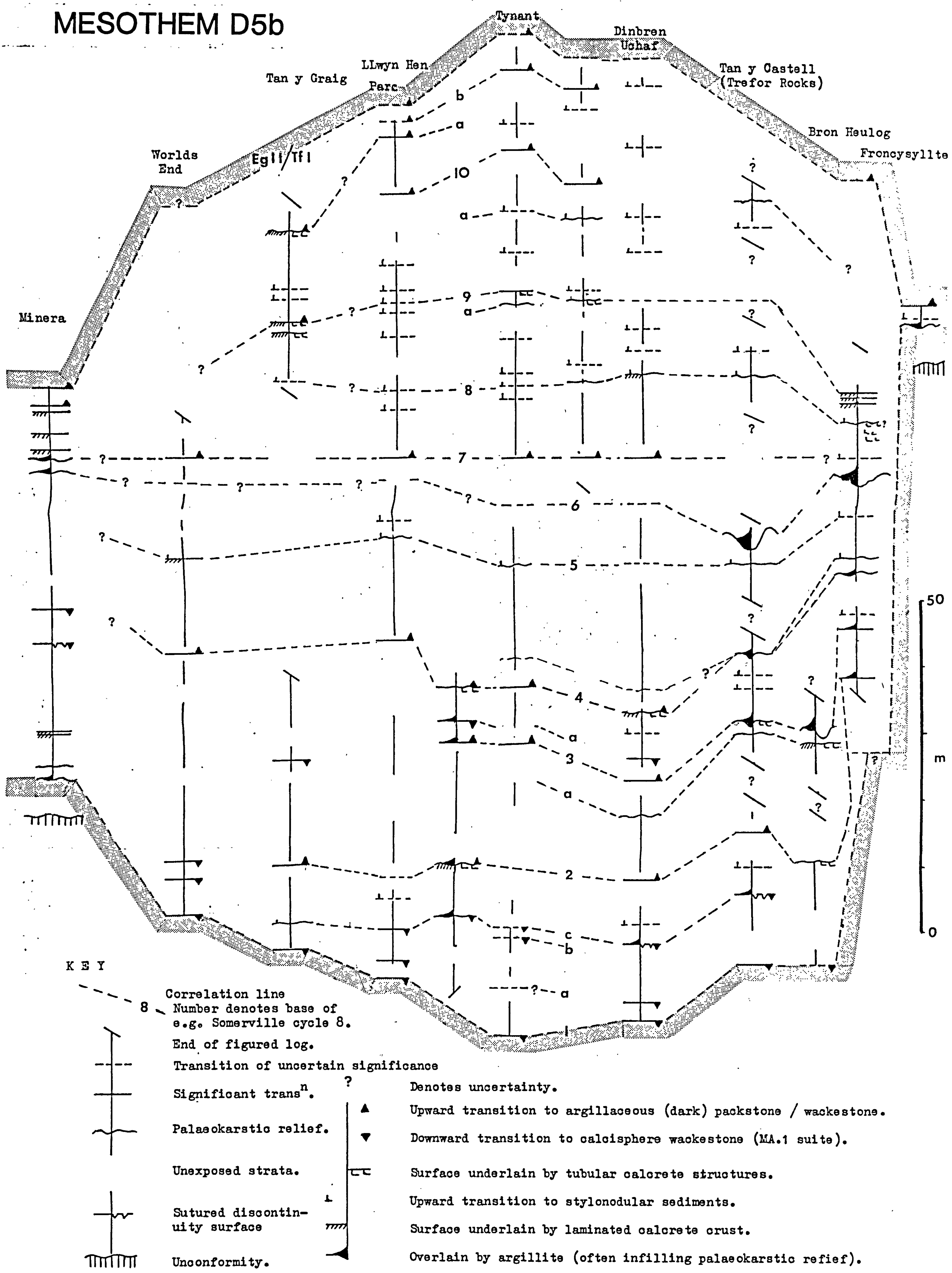
3.1. EGLWYSEG FORMATION

The Eglwyseg Formation was erected by Somerville (1977, 1979a) to include the Upper Asbian strata deposited within the 'confines' of the Llangollen Embayment (as defined herein), with its type section in the natural exposures of Tynant Ravine, on Eglwyseg Escarpment. The formation is conformable with both the underlying Tynant and overlying Trefor Formations. At its thickest development (around Tynant) it is 150m thick, but thins by onlap, southwards on to the Berwyns in the vicinity of Froncysyllte to zero, and northwards on to the Cyrn y Brain where it reaches a minimum of about 55m at Minera (see Chart B). North from Minera the formation thickens into the outcrops of the Vale of Clwyd, there termed the 'Loggerheads Formation' by Somerville (1977). Southwards, the Eglwyseg Formation is chronostratigraphically equivalent to the Llynclys Formation (see fig. 1 and section 3.2.).

Somerville (1977, 1979a) divided the Eglwyseg Formation into nine or ten 'minor cycles'. Ramsbottom and Somerville (in Somerville, 1979a, p. 337, 340) agreed that the term 'minor cycle' could readily be interchanged with 'cyclothem' (Ramsbottom, 1977) in the context of the Eglwyseg Formation. The term 'minor cycle' is here abbreviated to "cycle" unless otherwise clarified. It is used here in preference to 'cyclothem' (see section 1. 6.).

Somerville (1979a) discussed the detailed stratigraphy of his cycle scheme for the Eglwyseg Formation. Rather than repeat his

Fig 22 FORM AND DISTRIBUTION OF REGRESSION/TRANSGRESSION SURFACES : EGLWYSEG FORMATION



model for Eglwyseg Formation cycles, comprising 3 dominating phases:

phase C Thickly bedded or massive pale grey biosparite.
phase B Even bedded or wavy bedded dark grey biosparite.
phase A Calcareous shales, marls, rubbly micrites and biomicrites.

Problems arise in defining cycle boundaries when either clear-cut lithological transitions from Somervilles phase C to phase A are absent or subaerial emergence features (palaeokarstic surfaces, laminated calcrete crusts and clay palaeosols), are not present or not exposed.

Somerville's Eglwyseg Formation 'cycles 1 to 4' are readily defined by presence of these emergence features and phase C to A transitions, but in higher strata, limited exposure and apparently greater monotony of sediment types makes cycle identification less clear. The prolific development of 'rubbly micrites' (phase A) (termed 'stylonodular' here following Logan and Semeniuk (1976) in Wanless (1979)) interbedded with more massive bedded bioclastic peloidal grainstones (phase C) suggests that there may be more than the 14 minor cycles recognised therein.

The problem is one of degree. When does a cycle boundary become a minor fluctuation within a larger 'cycle'? (e.g.multiple cycle boundaries). This problem is also encountered in Mesothem D5a sediments, as indicated in section 2. 3. 1).

3.1.2.1. SEDIMENTS BETWEEN Eg. 1.AND Eg. 2. (Somerville's minor cycle 1).
Along the main outcrop of Eglwyseg the lowermost strata of the Eglwyseg Formation are well exposed. Somerville (1977, 1979a) recognised the presence of three thin calcisphere wackestone ('porcellanous

micrites') similar to the underlying Tynant Formation, but did neither include them within the Tynant Formation, nor give them full cycle boundary status.

One of these thin (≤ 30cm) beds at Dinbren Uchaf (Eg. 1c, fig. 22, Chart B) has a sutured discontinuity upper surface, indicative of intertidal exposure (see section 11.1.2.). They are interbedded with dark thin bedded argillaceous packstones (MA.2) and grainstones (MA.3), with rare gastropod-algal rudstone (MA. 4.1) lenses. The minor MA.1 phases represent short-lived regressive events within the first major transgressive phase of the Eglwyseg Formation.

3.1.2.2. SEDIMENTS BETWEEN Eg. 2.and Eg. 3. (Somerville's minor cycle 2)
The massive bedded 'phase C ' of this cycle is equivalent to the 'Basement Bed' of Morton (1879). Somerville (1977) noted that the well-developed laminated calcrete crust (Eg.2a) towards the top of the second minor cycle (G. R. SJ. 22854433) occurred 'up to 50cm below' the cycle top. Northwards from the Trefor exposure, this boundary is represented solely by a low amplitude (≤ 15cm)?palaeokarst, but does not correlate with Somerville's cycle 2 / 3 boundary at Tynant, (see fig.22 and Chart B). Somerville included the stylonodular and massive (4m) MA.3 unit that overlies the palaeokarstic / calcretised (Eg. 2a) surface along Trefor Rocks within his third minor cycle. This 4m unit, correlated with Somerville's third cycle at Tynant by Somerville (1977, 1979a) is a complete minor cycle along Trefor Rocks but has not been recognised north of Dinbren Uchaf (see fig.22). Along Trefor Rocks a very prominent palaeokarstic upper surface of this 4m cycle (Eg. 3) is infilled by maroon shales and conglomerates of Lower Palaeozoic slate and litharenite extraclasts. In interpreting this palaeokarst and conglomeratic infill as the minor cycle 3 / 4 boundary, Somerville neglected the presence of a fault, parallel to

the outcrop along Trefor, as depicted in fig. 23, that, at G. R. SJ 22854333, hides yet another minor cycle present between the conglomerate-overlain surface and the fourth minor cycle as defined by Somerville at Tynant (see fig. 23). The full succession, including this overlying minor cycle (eg.3. 4.) is exposed at G. R. SJ.22684350.

The minor fault cutting through the exposure at G.R. SJ22854333 believed by Somerville (1977) to indicate penecontemporary movement, is a small normal fault with a downthrow of ~ 1m eastwards, (fig. 23). The laminated shales and conglomerates infilling the deep palaeokarst (Eg.3) surface behaved in an incompetent manner adjacent to the fault plane, and do not represent primary sedimentation angles as interpreted by Somerville (1977, plate, 5.11).

3.1.2.3. <u>SEDIMENTS BETWEEN 'Eg 3' and 'Eg 4'</u> (Somerville's minor cycle 3)

Somerville (1977) recognised that this was one of the most stratigraphically variable of his ten cycles. 'Eg3a', occurring within Somerville's third cycle has a thinly developed calcisphere wackestone underlying it. This MA.1 phase, however, does not occur south of Dinbren Uchaf. On Trefor Rocks, the cycle is well exposed at G.R. SJ 22684350, with a palaeokarstic surface developed on its top (Eg 4) of up to 1m amplitude. This?correlates northwards to Tynant with a gently undulose palaeokarstic surface and underlying calcrete phenomena.

3.1.2.4. <u>SEDIMENTS BETWEEN 'Eg 4' and 'Eg. 5'</u> (Somerville's minor cycle 4)

Somerville's fourth minor cycle is very distinctive, including the "thick bed" of Morton. The transition from dark (bedded) fossiliferous grainstones (lithofacies L. 2) to massive pale regressive phase bioclastics, well exposed at Tynant and southwards (Plate 2, fig. E) may, however, be a microrelief palaeokarst ($\leq$ 10cm) (lacking calcretes). Its tenuous correlation is indicated in fig. 22.

STRATIGRAPHY OF MID-EGLWYSEG FORM'N , TREFOR

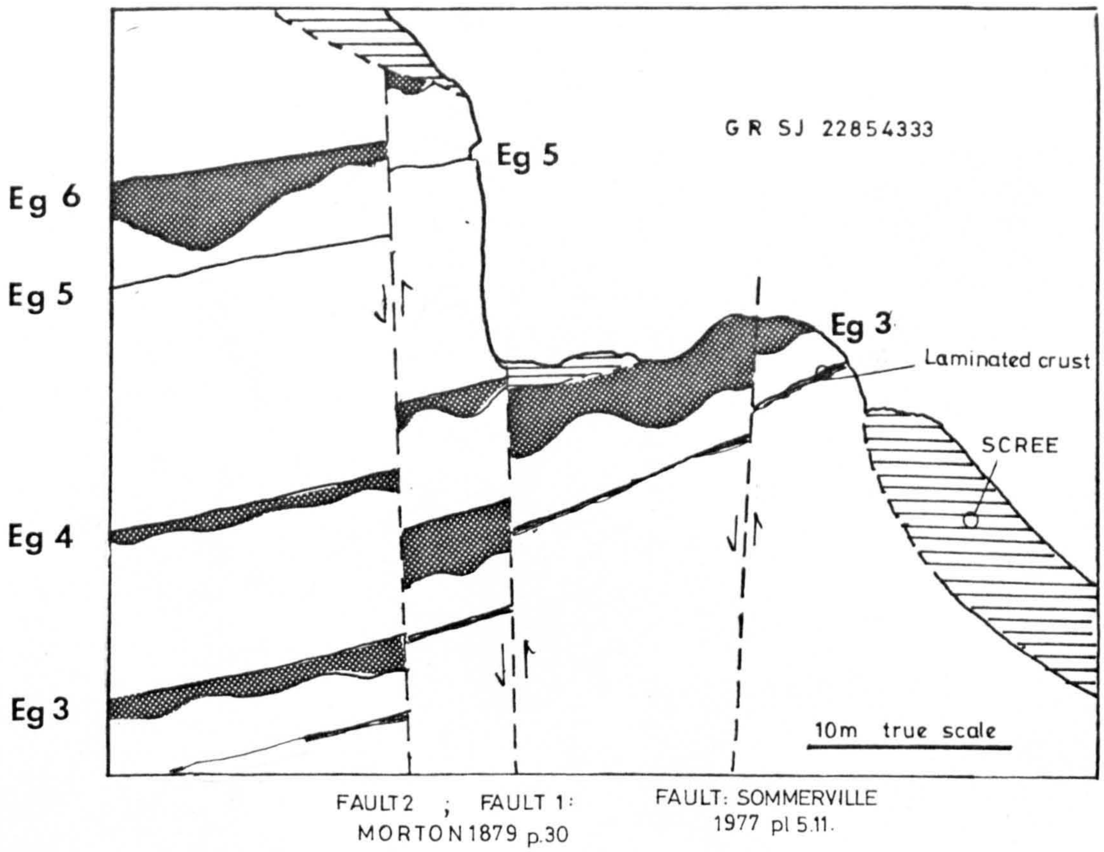


PLATE 2 Mesothem D5b.

A Bron Heulog Quarry, Eglwyseg Formation. Minor cycle boundaries 'Eg. 4' to 'Eg.9' marked. Note lensing nature of minor cycle 'Eg.5.6'. and compare with Plate 2, fig. C, G. R. SJ 238429

B Stylonodular phase at base of 'Eg 7.8'., Eglwyseg Formation, Bron Heulog Quarry. Note vague cm bedding and darker layers of maroon clay and maroon stained intercalations, G. R. SJ 23964284.

C Eglwyseg Formation, Trefor Rocks. Note minor cycle boundary 'Eg.6' with form C palaeokarstic relief (chapter 11). Contemporaneity of pot shown by maroon clay infill locally exposed, G. R. SJ 227434.

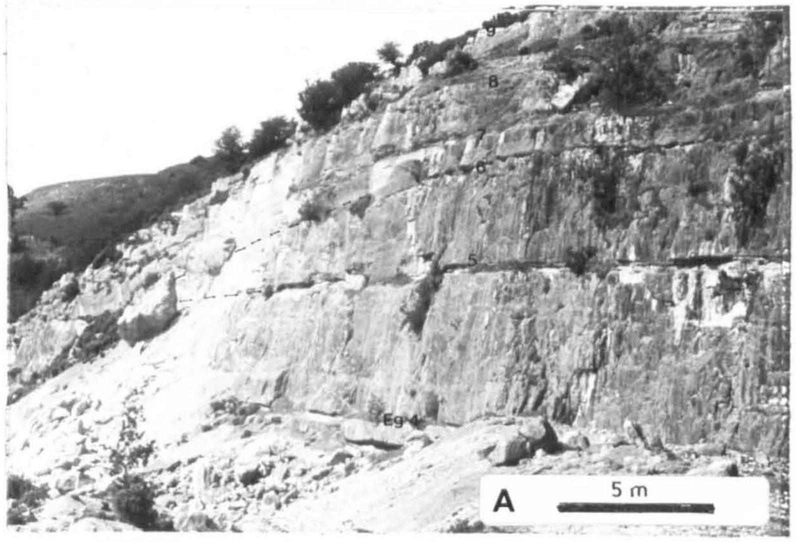
D Same feature as C, 100m south of C.

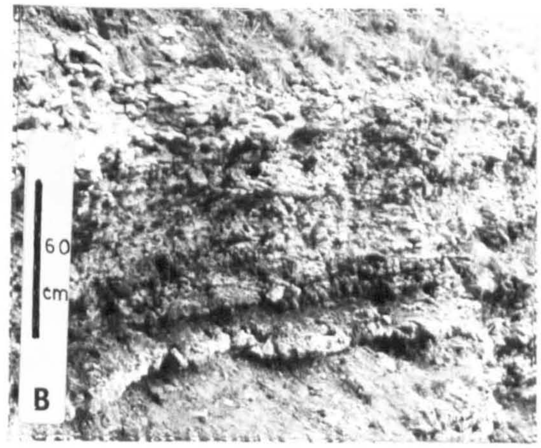
E Bedded dark biosparite (Lithofacies L.2) Eg. 4.5., Tynant. Note outlined coral colonies (L= lithostrotiontid; S= Syringoporoid) as laterally extensive and fragmented colonies. Uppermost surface seen in section may be palaeokarstic, although relief is $\lesssim$ 8cm, and calcrete structures are absent. G.R. SJ 22044549.

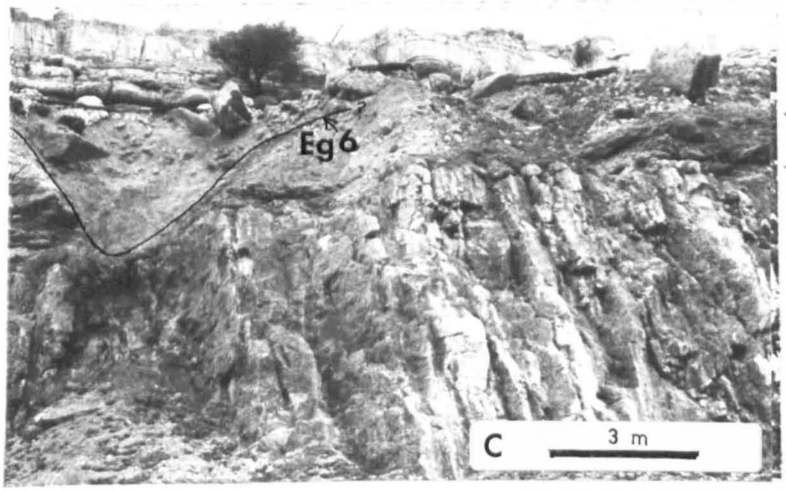
F Very thin bedded basal phase of minor cycle, Eg 8.9., Eglwyseg Formation, Llwyn Hen Parc, G. R. SJ.222463.

G Type section of upper strata of Llynclys Formation, Llynclys Hill, Porth y Waen. G.R. SJ 272239. Numbers signify minor cycle boundaries recognised herein.

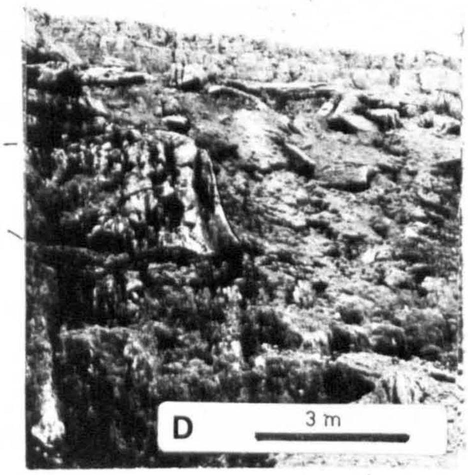
PLATE 2

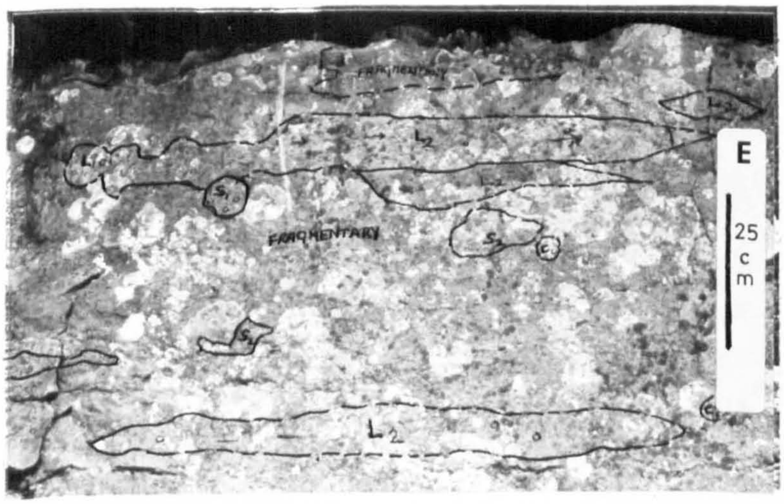


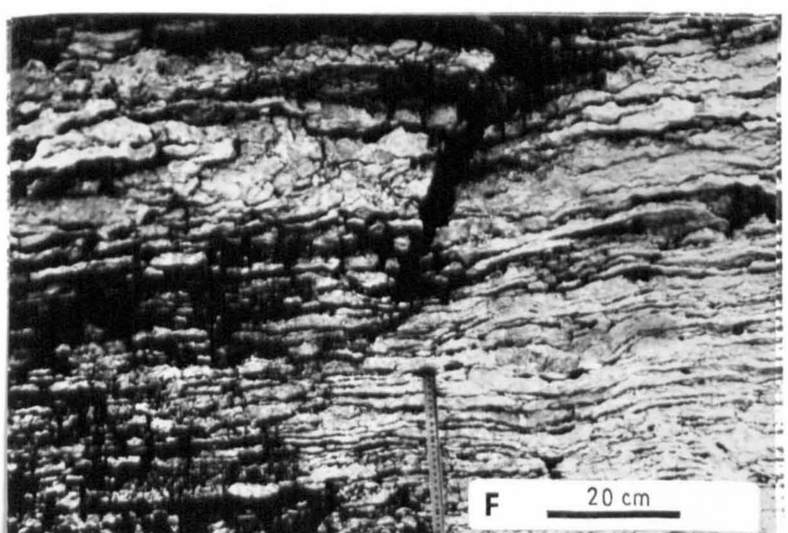




Eg 6
Eg5







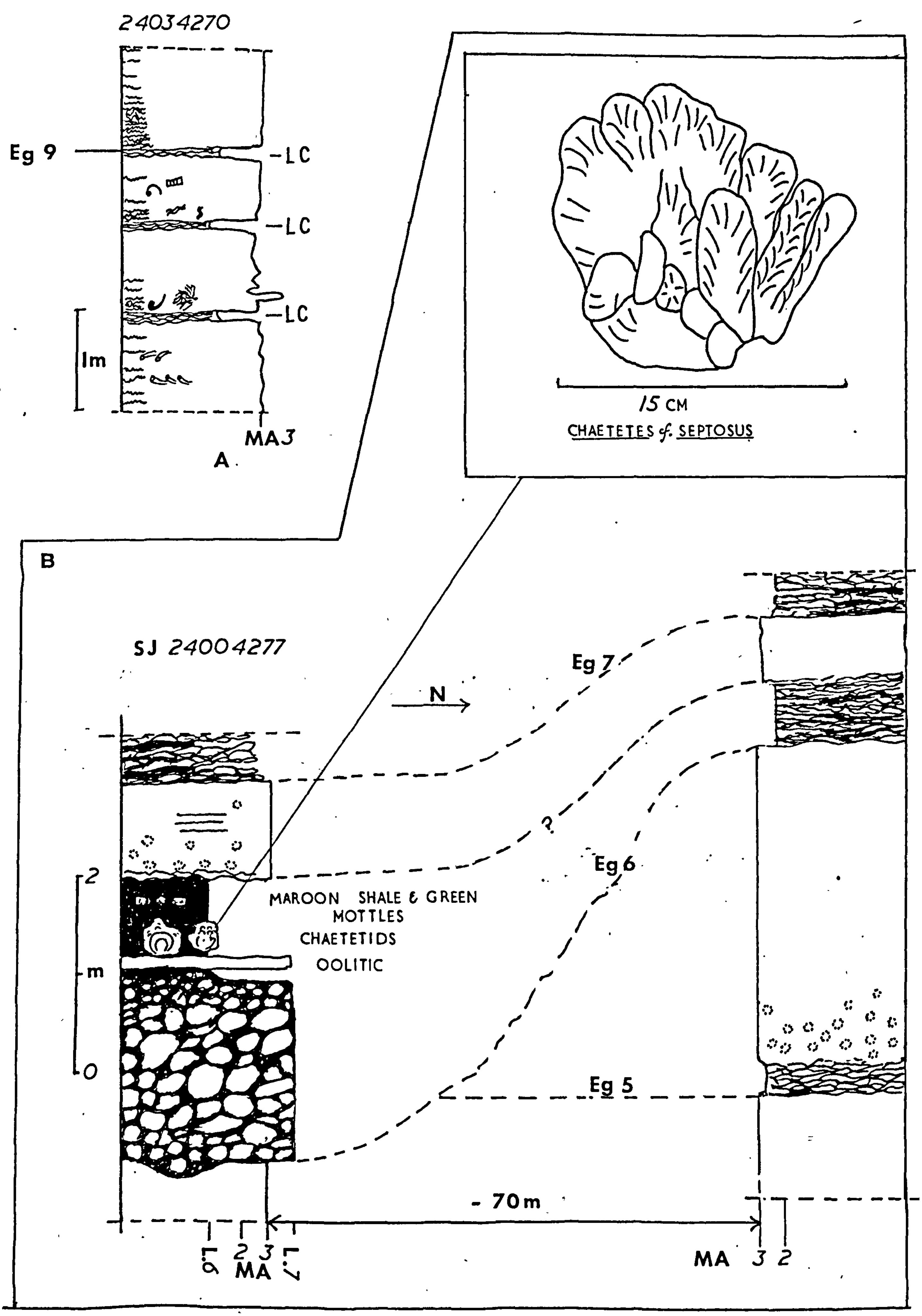
Ly9
Ly8
Ly7

3.1.2.5. SEDIMENTS BETWEEN 'Eg 5' and 'Eg 6' (Somerville's minor cycle 5)

Dominated by a thick massive MA.3 unit("24 foot" bed of Morton, 1879) this single cycle is readily correlated from Llwyn Hen Parc to Bron Heulog. At the southern end of Bron Heulog Quarry (G. R. SJ 24004277), a section exposing the sequence between cycles 4 and 7 of Somerville (1977) is well exposed, on the easterly wall of 'Fault 3' of Morton (1879, plate 4). Surface 'Eg 6' is deeply potted, in places down to the level of 'Eg 5' and its karstic relief infilled by a sequence of clast supported polygenetic conglomerates (including much Carboniferous material, maroon and cohre shales, and thin (5-10cm) impersistent carbonate lenses of intraclastic rudstones and conglomerates, with scattered well formed ferroan ooliths (Fig. 24b). These layers also contain in-growth position colonies of chaetetids, with tightly packed columnar growth morphologies (s/m. 2548, fig24b). Traced laterally northwards (c.80m) into Bron Heulog Quarry, a 0 - 5m lensing pale massive MA.3 horizon of cycle 'Eg 5. 6', is the lateral equivalent of the karst infill, indicated by the continuity of overlying strata (see Chart B). The lensing nature of this bed is readily visible within the quarry (Plate 2 fig. A). On Trefor Rocks (G.R. SJ 227434) the upper surface of this same massive pale MA.3 unit ('below' 'Eg 6') often obscured by scree and grass also shows prominent palaeokarstic relief (Chart B), there cutting down to near 'Eg 5' and infilled with maroon shales and extraformational conglomerates. The northward extent of this deep relief ($\leqslant$ 5m) palaeokarstic surface is not known, due to poor exposure, although its irregular upper surface, making contact with grass-covered scree extends to Tynant. This surface may provide a valuable correlation marker with other sequences of comparable age.(Plate 2, Fig. C & D)

Fig 24

ASBIAN EMERGENCE PROFILES, BRON HEULOG



3.1.2.6. SEDIMENTS BETWEEN 'Eg 6' and 'Eg 7' (Somerville's minor cycle 6)

This cycle, as defined by Somerville (1977) is dominated by stylonodular packstones and grainstones, with thin shale intercalae. (see section 3.1.3.1).

3.1.2.7. SEDIMENTS BETWEEN 'Eg 7'and 'Eg 8' (Somerville's minor cycle 7)

Somerville's cycle 7 does not contain any further indicators of minor cyclic developments within it, although transitions from massive to stylonodular sediments occur within the upper phase of the cycle. These are not believed significant, as the stylonodular MA.3 lacks intercalated clay, that is a more significant transgressive indicator, than the stylonodularity itself.

At Bron Heulog the base of the cycle is represented by a 2m argillaceous stylonodular phase, whilst its upper 'boundary Eg8' is palaeokarstic, of low relief, and with a poorly developed laminated calcrete crust at Dinbren Uchaf.

3.1.2.8. SEDIMENTS BETWEEN 'Eg 8' and 'Eg 9' (Somerville's minor cycle 8)

' Eg.9' is a master bedding plane (sensu Schwarzacher, 1958) with a poorly developed laminated calcrete crust, occasionally only visible as tubular calcrete structures (see section 11.3.1.3). Not all sections show palaeokarstic relief, and the flatness, but irregular calcretisation of the surface at Tynant and Tan y Graig suggest that a degree of predepositional scour occurred on the succeeding transgression.

Beneath this surface are other closely spaced emergence surfaces with palaeokarstic relief and or calcretisation phenomena.

At Bron Heulog, the upper half of Somerville's cycle 7 has here been placed between 'Eg 8' and 'Eg 9', and therefore, correlated as equivalent to Somerville's cycle 8 at Tynant. This correlation is due to the similarity of the multiple emergence events immediately below boundary 'Eg 9': Beneath 'Eg 9' at Bron Heulog are 'Eg 8a' and 'Eg 8b', each defined by a prominent smooth, slightly undulose (5cm) bedding plane, underlain by a laminated calcrete crust (fig.24a). These 3 laminated calcrete crusts are separated respectively by 90 and 70cm beds of MA.3 sediments, containing occasional reworked crust clasts up to 8cm long. A 'double' crust (separated by 60cm of MA.3, containing a Chaetetes Syringopora biostrome) is also exposed at a (?) similar level at Tan y Graig. This situation is very similar to that observed in the upper strata of the Eglwyseg Formation at Minera (see chart D), however, in this latter section, the crusts are not associated with any other emergence indicators (no palaeokarstic surfaces, nor lithological transitions), and they are locally removed by pressure solution. Whether or not these units between multiple emergence features represent discrete cycles is discussed in chapter 12.

3.1.2.9. SEDIMENTS BETWEEN 'Eg 9' and 'Eg 10' (Somerville's minor cycle 9)

Somerville (1977) recognised an argillaceous stylonodular phase 2 to 3m below his ?cycle 9/10 boundary. This stylonodular phase, well exposed at Tynant (G. R. SJ 22424534) has a low amplitude palaeokarst underlying it, ('Eg 9a'). This surface is laterally equivalent to a massive MA.3/ stylonodular transition, either unexposed or without any apparent palaeokarstic relief.

'Eg 10', of uncertain form and position according to Somerville, (1977) and possibly defined by a laminated calcrete crust above Trefor Rocks (G. R. SJ 22924338), is poorly exposed in sections along the upper crags of Eglwyseg, comprising a transition to thin-bedded argillaceous packstones (Somerville's phase B2). Its contact with the underlying massive and stylonodular MA.3 is not exposed.

3.1.2.10 SEDIMENTS ABOVE 'Eg 10'. Somerville's minor cycle 10)

Chart B shows probably two further boundaries ('Eg 10a' and 'Eg 10b') above 'Eg 10', recognised by transitions to thin bedded argillaceous packstones, each defining minor cycles 2 to 6m thick. Poor exposure suggests that other cycle boundaries are possibly hidden.

3.1.2.11. GENERAL CORRELATION PROBLEMS.

1) The Minera succession, according to Somerville, comprises Eglwyseg Formation minor cycles 1 - 6, with higher strata absent. His correlations are tenuous, using visible emergence phenomena, stylonodular phases, and not, as in the case of the underlying Tynant Formation the presence of regressive MA.1 lithologies as cycle upper phase markers, but which are present in the Eglwyseg Formation at Minera. Charts B and E show the Minera sequence as logged herein, recognising nine emergence horizons from palaeokarsts and calcrete laminated crusts alone. Sutured discontinuity surfaces (section 11. 1) represent brief emergence events , and MA.1 suites (especially fenestrated) represent regressive peritidal phases. Using their presence to predict cycle boundaries, in excess of 13 'cycles' occur in the Minera sequence. (Many of the thinner MA.1 phases, however, disappear between sections taken 100m apart). There may be further emergence features still unrecognised in the more poorly exposed parts of the section.

The correlation of particular units with their more southerly counterparts is very subjective in this 'condensed sequence' at Minera. Two prominent palaeokarstic surfaces in the Minera outcrop are tentatively correlated with those on top of the fifth and sixth Somerville cycle on Eglwyseg (see fig 22) i.e. 'Eg 6' and Eg.7', but the lack of distinctive lithologic transitions and faunal horizons renders even this correlation conjectural. The deeper (form 'C ') palaeokarst features may span whole cycles of deposition.

The uppermost cycles may be absent, as suggested by Somerville (1979a) reflecting the overall regression towards the top of Mesothem D5b (Ramsbottom 1977) as indicated in the Llynclys Formation at Oswestry. (see on).

The general development of the lower 25m of Eglwyseg Formation at Minera is markedly different to that within the embayment 'depocentre' 7km to the south, with a greater proliferation of MA.3 sediments of shoaling high energy facies (Chart E) reflecting its location atop the positive Cyrn y Brain axis, the principle cause for the attenuated sequence.

2) At Froncysyllte (G. R. SJ26984199) the Eglwyseg Formation is chronostratigraphically represented by 4m of strata, resting above an erosion surface and coarse clast supported conglomerate of the top regressive MA.1 unit of the Tynant Formation(?) (Fig.40). This 4m is also chronostratigraphically comparable to the Llynclys Formation. It comprises a single cycle with a basal 2m pale stylonodular and argillaceous MA.2 phase overlain by a massive MA.3 unit. The transition to argillaceous thin bedded packstone/wackestone of the overlying Trefor Formation is apparently a planar surface, without evidence of subaerial emergence. Whether the 4m represents the uppermost Eglwyseg Formation cycle as exposed at Tynant is unclear. The regression associated with the top of Mesothem D5b (Ramsbottom 1977) suggests that it is unlikely to represent the uppermost cycle, however (cf section 3.2.2.1.)

3.1.3. Cycle Form.

Somerville (1977) recognised the Eglwyseg Cycles as shoaling deposits, with a basal (phase A and B′ of Somerville 1979a) transgressive unit, and an upper 'shallowing' unit (phase 'C ') terminated by subaerial emergence. This sequence was likened by Somerville (1977) to the cycle model of Ramsbottom (1973), with sedimentation in response to sea-level fluctuations with a possible eustatic control. Chapter 9 analyses this concept from a microfacies viewpoint.

The 'ideal' cycle†, however, as depicted by Somerville (1979a, fig. 2) is very variable in thickness. The relative thicknesses of its constituent phases vary between individual cycles, from those dominated by the massive shoaling (MA.3) deposits of Somerville's 'phase C' (e.g. Cycle 'Eg 5. 6', as defined here) to those with thickest basal 'transgressive' phases (e.g. cycle 'Eg 6. 7'). Many of the lower cycles developed at Minera and within Somerville's first minor cycle along Eglwyseg contain MA.1 phases, therefore showing similarities to the Tynant and Whitehaven Formations.

3.1.3.1. TRANSGRESSIVE PHASES. The basal 'transgressive phase' is dominated by either MA.2 sediments or an intimate interlayering of MA.3 and clay intercalae, the latter producing the stylonodular texture referred to as 'AO' by Somerville (1977)

MA.2 (ARGILLACEOUS WACKESTONE-PACKSTONE) DOMINATED BASAL PHASES.

Their presence towards cycle bases, close to the 'master bedding planes' of the cycle boundaries leads to sporadic, and often poor exposure. Somerville (1979, fig. 1) grouped these sediments into a 'phase B', dividing them on lithological make-up into thick-bedded pale grey poorly fossiliferous micrites (B1): even bedded, dark grey, highly fossiliferous biosparites (B2); thin and wavy bedded dark grey biosparites (B3).

Basal phases of dark argillaceous packstones (MA.2) are well bedded L. 3 (5 to 50cm) with shale partings and occasionally underlain by fossiliferous calcareous shale that may, in turn, rest up on a palaeosol and/ or palaeokarst. They often underlie Somerville's

† cf. fig. 27a.

B 2 (Lithofacies L.2). Traced both northwards and (Fig. 27) southwards from the type section at Tynant these phases generally become thinner or absent on to the Cyrn y Brain and Berwyn axes. A similar relationship between L2 & 3 sediments and underlying tectonic elements is found in the Trefor Formation (see section 5.1.4.).

These L.2 sediments at times support a diverse coral assemblage, with whole and fragmentary colonial and solitary corals locally accounting for up to 20% of the sediment (e.g. cycle 'Eg 3. 4', Trefor Rocks, G. R. SJ 22684350 (see section 3.1.2.4) and 'Eg 4. 5' Tynant (G. R. SJ 22044549)). Overlying MA.3 sediments do, however, support an equally diverse fauna, but not in this exceptional abundance.

STYLONODULAR AND ARGILLACEOUS MA2 /MA.3 BASAL PHASES.

Somerville (1977) recognised these sediments as transgressive deposits overlying emergence surfaces. They are characteristically either very thin bedded† ($\ll$ 15cm) pale or stained wackestones to grainstones, with irregular millimetre to cm thick grey, green or maroon clay intercalae, or are irregularly stylonodular†. In petrographic make-up they are equivalent to overlying massive MA.3 sediments, although the presence of the clay has often induced masking neomorphic fabrics (Longman, 1977) (see section 9. 1. 2), and probably enhanced the stylolitisation process. Noteably present above 'Eg 5', 'Eg 6', 'Eg 7' and 'Eg 8', these sediments support a similar and as variably developed coral sponge brachiopod assemblage as overlying massive phases. Stylonodular sediments in upper phases lack shale intercalations.

3.1.3.2. UPPER 'REGRESSIVE' PHASES. The junction between basal L.2 & L.3 and upper massive-bedded L . 1 sediments often grades through a 'pseudobreccia' mottled zone, with darker irregular ½ to 5cm sized mottles set in a paler matrix (section 9. 2. 2).

† Plate 2, fig. F † Plate 2, fig. B

LATERAL VARIATION IN MINOR CYCLE TRANSGRESSIVE PHASE THICKNESS
(LITHOFACIES L.2 + L.3 + L.6 EGLWYSEG FORMATION)

(in metres)

	NORTH Minera	World's End	Tan y Graig	Llwyn Hen Parc	Tynant	Dinbren Uchaf	Tan y Castell	SOUTH Bron Heulog
Eg10b		?	?	<1.5	>0.5	>0.5	?	?
Eg10a	0	?	?	0.8	?	?	?	?
Eg10	0	?	?	>1.5	>2.0	>0.5	?	?
Eg 9	0	?	0.5	?	~2.0	?	?	?
Eg 7	0	?	?	6.0	2 to6	2 to 6	?	< 2.0
Eg 4	0	2 to 4	<2.5	> 2.0	3 to 6	6.0	2.0	0.5
Eg 3	0	<2.0	< 2.0	?	> 3.0	>3.0	2.0	~0.4
Eg 2	0	< 4.0	<5.0	~3.0	< 6.0	<7.0	<6.0	~ 6.0
Eg 1	0.5	8.0	4.0	8.0	13.0	<12.0	<12.0	?

↑ Associated underlying reference cycle boundary.

Fig 27c

The massive bedded MA.3 phase comprises homogenous to parallel and cross laminated grain supported sediments. Detailed logging shows that pseudobreccia mottling is pervasive through these units, but is mostly vague, and often concentrated in 10 to 30cm 'beds'. Smaller irregular mottles locally define vague centimetre lamination. Slightly darker 5 to 40cm layers of matrix free coarser grainstones can be traced along good outcrop for 100 s of metres (e.g. cycle 'Eg. 4. 5', Trefor Rocks) (fig.25), and reflect a minor internal rhythmic sedimentation pattern.

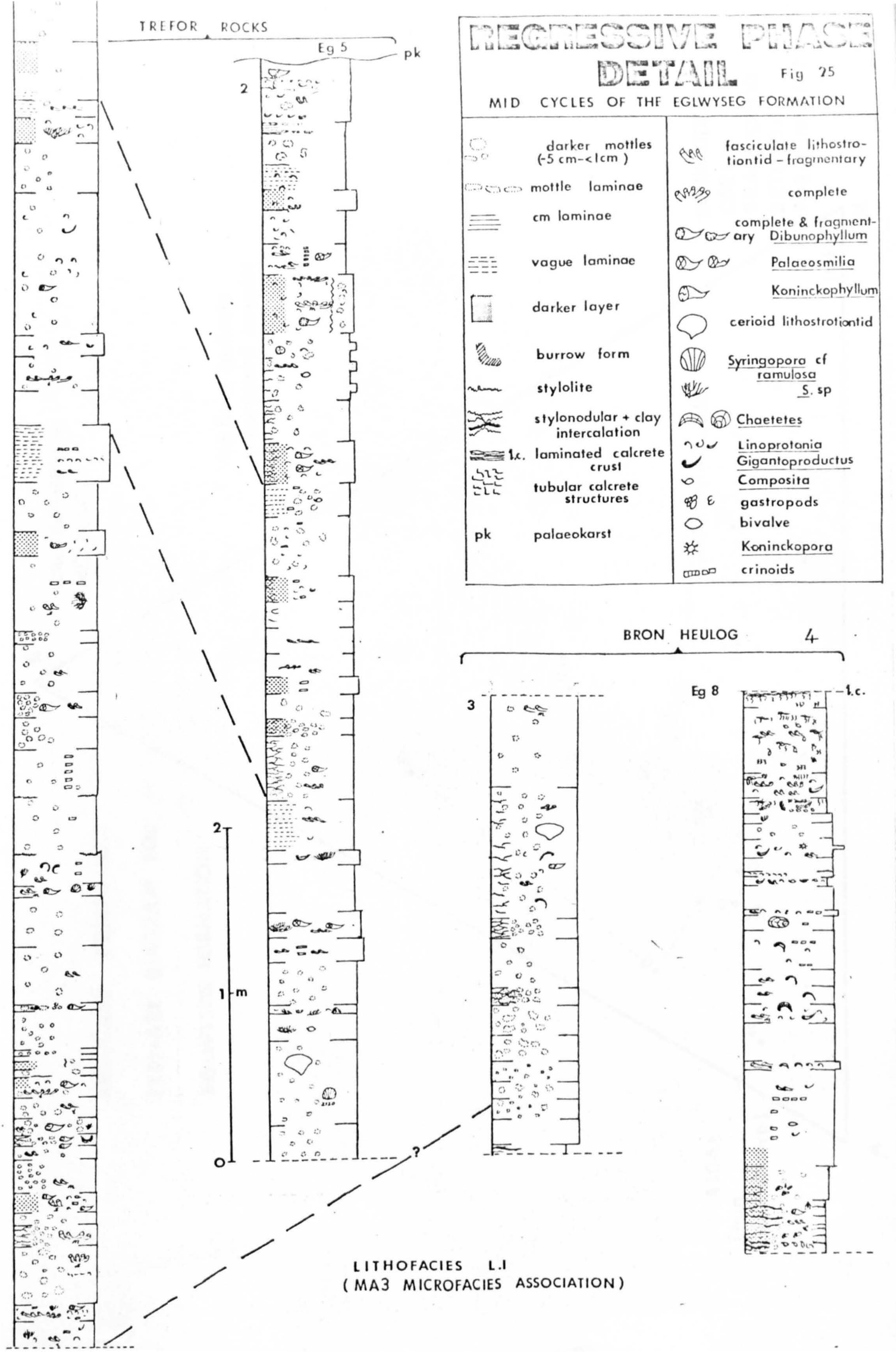
Only rarely are MA.1 phases developed as regressive facies below cycle boundaries. They are especially common within the basal cycle 'Eg. 1.2' along Eglwyseg, and within the cycles below ?'Eg.6' at Minera.

3.1.3.3. <u>CYCLE TRENDS.</u> As with the Tynant and Whitehaven Formation, bioclast compositions of the Eglwyseg Formation sediments are plotted on the tenary C-A-M diagram (figs.26 &17). The plot of their distributions shows similar trends to these other formations, but two prime differences arise:

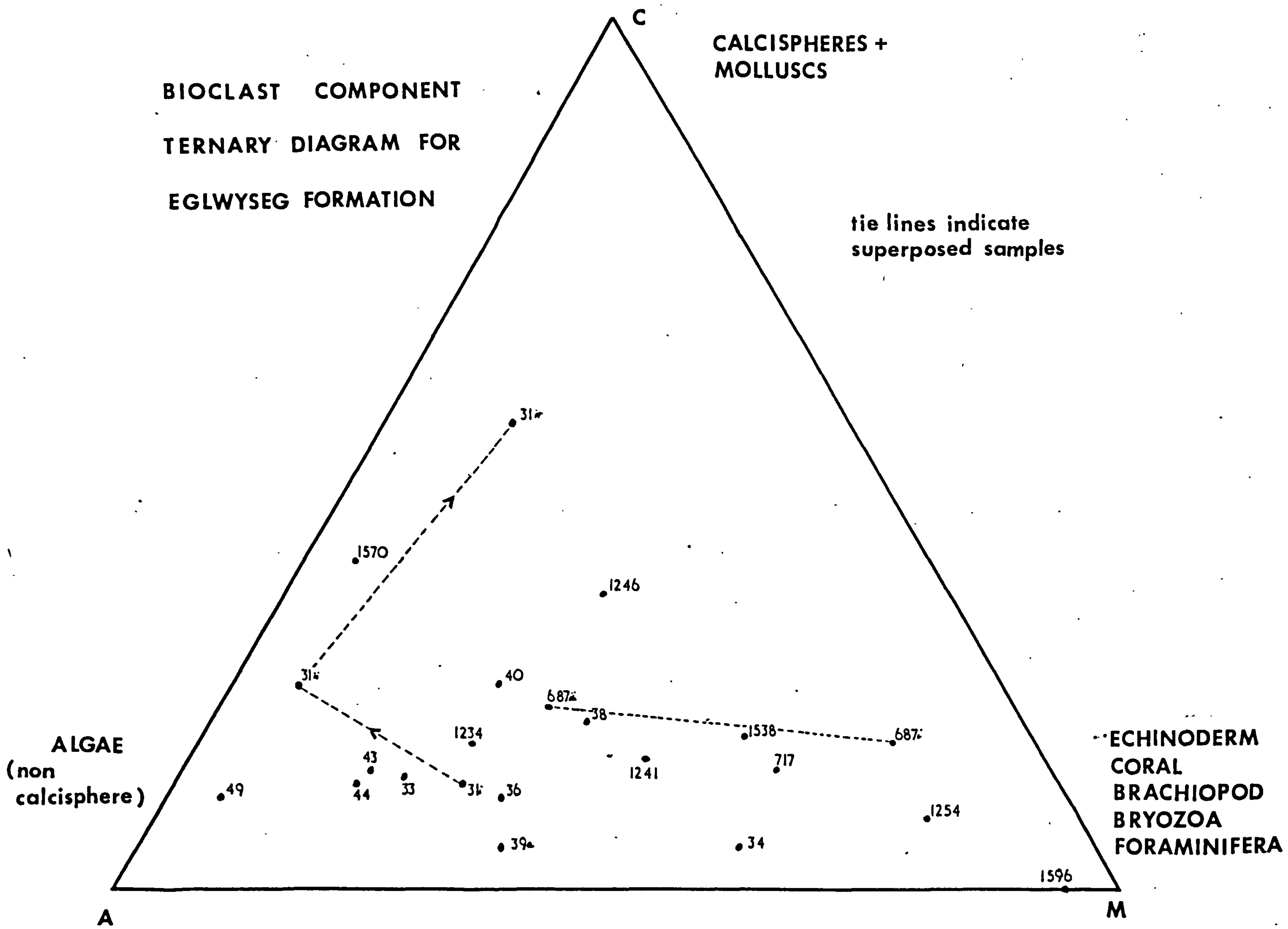
1) The Eglwyseg Formation has a poorly defined composition 'tail' towards the C end member, due principally to the relative lack of MA.1 lithologies, and masked by '(2)'.
2) There is a greater spread of MA.3 and MA.2 bioclastic compositions than apparent in the other Formations investigated, producing a broad central field to the C A M diagram.

3.1.4. <u>Palaeontology.</u>

The Eglwyseg Formation contains a typical Asbian fauna, with <u>Palaeosmilia murchisoni</u> (Milne Edwards and Haime), <u>Dibunophyllum</u>

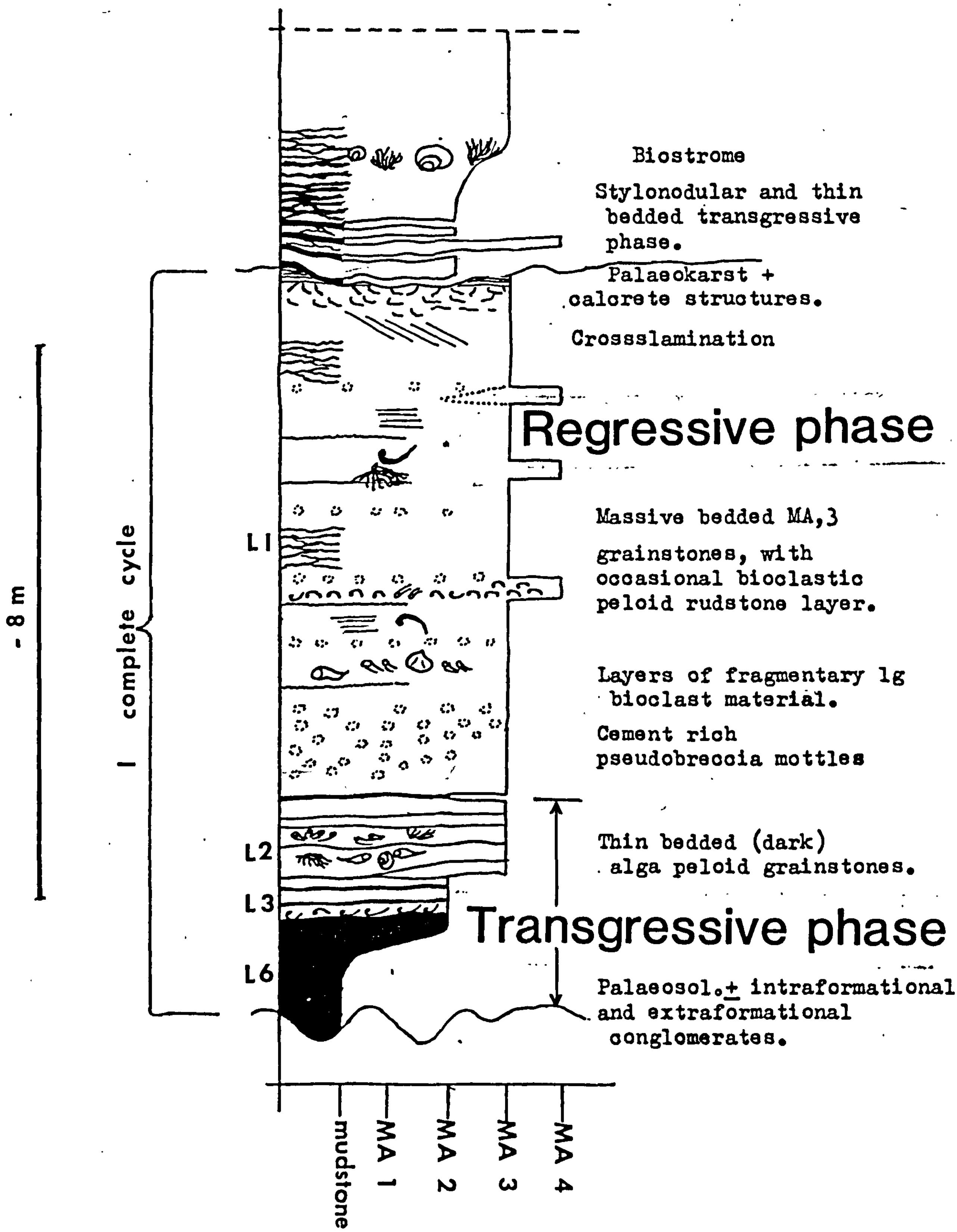


REGRESSIVE PHASE DETAIL
Fig 25
MID CYCLES OF THE EGLWYSEG FORMATION
darker mottles (-5 cm-<1cm)
mottle laminae
cm laminae
vague laminae
darker layer
burrow form
stylolite
stylonodular + clay intercalation
l.c. laminated calcrete crust
tubular calcrete structures
pk palaeokarst
fasciculate lithostrotiontid - fragmentary
complete
complete & fragmentary Dibunophyllum
Palaeosmilia
Koninckophyllum
cerioid lithostrotiontid
Syringopora cf ramulosa
S. sp
Chaetetes
Linoprotonia
Gigantoproductus
Composita
gastropods
bivalve
Koninckopora
crinoids
TREFOR ROCKS
Eg 5
pk
2
2
1 m
0
?
BRON HEULOG
4
3
Eg 8
l.c.
LITHOFACIES L.I
(MA3 MICROFACIES ASSOCIATION)


BIOCLAST COMPONENT
TERNARY DIAGRAM FOR
EGLWYSEG FORMATION
C
CALCISPHERES +
MOLLUSCS
tie lines indicate
superposed samples
ALGAE
(non
calcisphere)
A
M
ECHINODERM
CORAL
BRACHIOPOD
BRYOZOA
FORAMINIFERA
1570
1246
40
1234
43
44
33
36
49
1241
717
1538
38
1254
34
1596
fig 26

COMPOSITE SEQUENCE MINOR CYCLE

EGLWYSEG FORMATION



bourtonense (Garwood and Goodyear), Linoprotonia hemisphaerica (J. Sowerby), Lithostrotion sp. nov. [J. R. Nudds in fide], Lithostrotion sociale (Phillips) L. decipiens (McCoy) L. junceum (Fleming), Daviesiella sp., Gigantoproductus cf maximus

Davidsonina septosa (Phillips), a characteristic late Asbian brachiopod in the north of England and Derbyshire, has not yet been recorded from North Wales.

Appendix IV lists the faunas recorded in the Eglwyseg Formation from collection and observation in this study.

3.2. LLYNCLYS FORMATION (New Term)

Derivation of name: From Llynclys Hill, 1km SW of Llynclys, G. R. SJ 2723, upon which the type section outcrops.

Type localities: Lower strata of the Llynclys Formation: upper 32m of Nant Mawr Quarry, G. R. SJ 252250; upper strata of the Llynclys Formation; disused quarry, Llynclys Hill, G. R. SJ 272239 (Plate 2 fig. F).

Lithostratigraphic definition: The base of this formation, and junction with the top of the underlying Whitehaven Formation is well exposed in the east face of Nant Mawr Quarry (G. R. SJ 25242500) defined by a prominent "Form A" (chapter 11) palaeokarst, overlain by a green and black clay, (palaeosol) 32m from the quarry top (chart C). The formation comprises cycles of similar basic form to the Eglwyseg Formation, its chronostratigraphic counterpart, with basal thin bedded maroon stained MA.2 and MA.3 units overlain by massive or stylonodular MA.3 phases. Cycle boundaries are defined by either palaeokarstic surfaces or lithological transitions. The formation includes both Morton's (1879) '40 inch bed', '30 feet bed' and the overlying "Rubbly Beds" of Wedd et al (1929). There is no evidence to support a prolonged non-sequence at the base of the Llynclys Formation, of Asbian / Holkerian boundaries in the North of England.

The upper boundary with the Trefor Formation is unexposed, but is defined as the first transition to the dark argillaceous and thin bedded packstones of the Trefor Formation basal unit, as exposed at the base of Dolgoch, Pant Hir, and Treflach Quarries (see fig.30).

Thickness and Distribution.

The sequence is always incompletely exposed. The type section at Llynclys Hill exposes an upper 17m of the formation only. Around Whitehaven/Llynclys, the Llynclys Formation is at its maximum development of about 40m. Northward it thins by?onlap followed by offlap,to zero between Llawnt and Craig y Rhiw (a distance of 8km) i.e. the lowermost cycle(s) onlap the Whitehaven Formation and overstep the Berwyns, whilst upper cycles of the Llynclys Formation offlap earlier cycles to an apparent depocentre near Llynclys.

Stratigraphic Nomenclature:

The stratigraphic numbering scheme conforms to that used throughout this thesis. All boundaries are suffixed 'Ly' and numbered in stratigraphically 'younging' order (see section 3. 1. 1.)

3.2.1. The Lower Cycles

The Llynclys Formation is informally divided into a lower and upper division along 'Ly 3'.

3.2.1.1. <u>SUCCESSION BETWEEN 'Ly 1' AND 'Ly 2'</u> - The first minor cycle - Best exposed in Nant Mawr Quarry, this cycle is $\leqslant$5m thick, bounded at base by a form 'A' palaeokarst (50cm max. relief) with infilling black and green clays. Northwards at Craig Sychtyn, this palaeokarst also has a 50cm relief ., whilst at Craig y Rhiw it is a near planar form 'D' surface covered by a thin maroon shale parting (see Chart C).

A thin basal 40cm MA.1 phase is developed, with fenestral fabrics and with an overlying S. D. surface†, possibly comparable to the thin MA.1 phases present in the transgressive phase of the first Eglwyseg Formation minor cycle, and indicating a gradual transition to cycles of Llynclys Formation form from Sychtyn Member cycles.

The succeeding massive and mottled MA.3 unit includes the "40 inch" bed of Wedd <u>et al</u> (1929, p.109), with a stylonodular basal phase, succeeded by coarse laminated grainstones towards the top of its $\leqslant$ 3½m at Nant Mawr. At Craig Sychtyn it is represented by 2m of stylonodular, overlain by 4m of massive, MA.3 sediments, whilst southwards in the Whitehaven quarries this horizon has been preferentially dolomitised, masking its original fabric, although the overlying ('Ly 2') Form B palaeokarst is clearly visible. 'Ly 2', traced northwards, becomes a prominent 'Form A†' surface', that also defines a prominent bench towards the top of Craig y Rhiw (G. R. SJ 23852980). At Nant Mawr 'Ly 2' is developed on a thin $\leqslant$80cm multiple MA.1 phase, which is separated by a low relief S. D. surface†, with 20cm of MA.3 grainstones and an intraformational conglomerate horizon.

3.2.1.2. <u>SUCCESSION BETWEEN 'Ly 2' AND 'Ly. 3'</u>. This cycle is dominated by the '30 feet' bed of Morton (1879). Where its lower horizons are exposed they comprise 1½ to 2m of thin bedded (5 -20cm) maroon stained MA.3 sediments with maroon shale partings, and a locally well developed

† Chapter.11

stylonodular form. At Whitehaven Quarry (G. R.SJ 26452450), 60cm of MA.1 sediments shows that this cycle also retains transgressive features.

The overlying massive MA.3 phase is 6½m thick at Nant Mawr, the only locality where it is fully exposed, although it forms a scarp all along outcrop to the northern end of Craig y Rhiw Crags (G. R. SJ 24002995). Its character is similar to the massive MA.3 developments of the Eglwyseg Formation (especially below 'Eg 5' and 'Eg. 6'), with a prominent mottled basal 2m. 'Ly 3' is a transition to stylonodular sediments with maroon shale intercalae. The surface is obscured by staining and stylolite weathering in the underlying sediment.

3.2.2. <u>The Upper Cycles.</u>

3.2.2.1. <u>SUCCESSION ABOVE 'Ly. 3'</u>. Above 'Ly 3' the general character of the succession changes due to the presence of maroon stained stylonodular sediments and thin bedded basal phases with maroon shale intercalae, accounting for about 50% of the sequence. ('The Rubbly Beds' of Wedd <u>et al</u>. (1929) are exposed along the north and east slopes of Llynclys Hill, and their base at the very top of Nant Mawr Quarry only. They are probably 25m thick at Llynclys/Whitehaven but 8 - 10m at the southern end of Craig y Rhiw where they mark a vale beneath the Trefor Formation escarpment (G. R. SJ. 236295). This northwards thinning may be accounted for by a persistent off-lap of these upper cycles, off the Berwyn Axis.

Precise cycle boundaries in these strata are not well defined. There is a lack of prominent palaeokarstic relief, and upper sediments of underlying cycles are maroon stained with haematite along stylolites, therefore appearing to grade into the stylonodular-haematitic argillaceous basal phase of the above cycle (Plate 2 , fig G ;

Chart D.) These upper cycles are $\leq$ 6m thick, (most about 4m) and at least 7 are present, but more may be obscured due to lack of continuity between sections at Llynclys and Nant Mawr. The base of cycle 'Ly 8. 9' (Chart C) has a Gigantoproductus cf maximus shell bed, succeeded by 2m of nodular calcareous shales and maroon stained thin bedded MA.3 lithologies.

3.2.3. Comparisons of Llynclys Formation Lithostratigraphy.

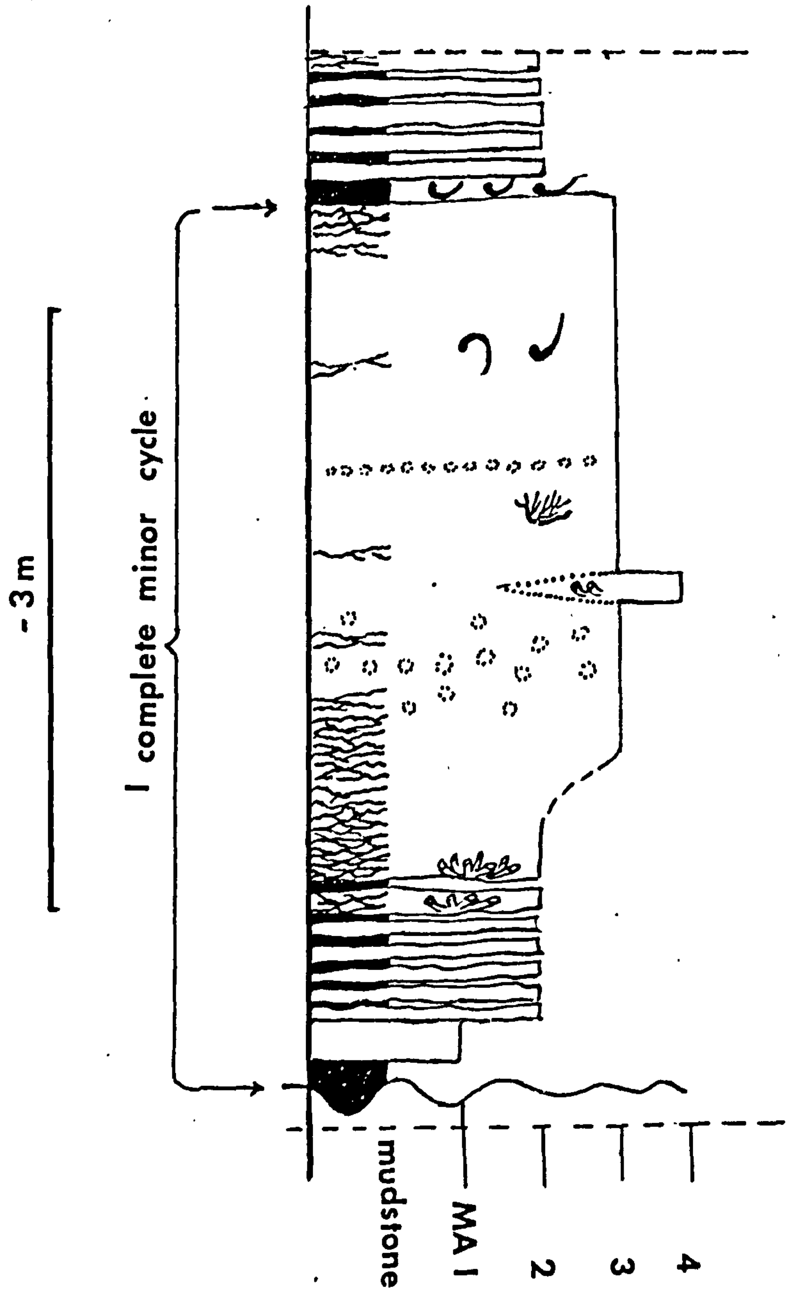
Morton (1879 p.109) correlated his equivalent of the upper cycles of the Eglwyseg Formation with the Llynclys Formation, due to both their equivalent positions beneath the Trefor Formation (Upper Grey Limestone) and their lithological similarities. Wedd et al (1929) correlated the Llynclys Formation equivalent with the whole Eglwyseg Formation equivalent, as followed here (fig. 1).

Fig 27b depicts a composite sequence Llynclys Formation cycle. The Llynclys Formation basal cycles lack the thick basal thin bedded MA.2/MA.3 transgressive phases common in the lower cycles of the Eglwyseg Formation. However, as shown in section 3. 1. 3, these basal phases of the Eglwyseg Formation cycles vary considerably within the confines of the Llangollen embayment, in general thinning away from the embayment centre. The absence of thick MA.2 basal phases in the Llynclys Formation does not preclude the contemporaneity with either the higher or lower Eglwyseg Formation cycles.

Morton (1879) believed that the "Thick Bed" ('Eg. 4. 5') of Llangollen was the lateral equivalent of the "30 feet bed" ('Ly 2. 3'). This is a very probable correlation due to the prominence (?and probable great lateral extension) of the 'Eg. 4. 5' cycle. This correlation would, however, imply that 'Eg. 1' to 'Eg. 4' are represented by 'Ly. 1. 2' and its bounding palaeokarsts.

COMPOSITE SEQUENCE LLYNCLYS FORMATION MINOR CYCLE

fig 27b



3.2.4. Llynolys Formation Biostratigraphy.

The degree and quality of present exposure neither facilitate faunal observations nor collection. Wedd et. al (1929) noted the presence of Palaeosmilia murchisoni, and appendix IV lists the fossils identified in this study. It has a typically Asbian suite including Dibunophyllum bourtonense, but there is insufficient evidence available to determine which cycles of the Llynolys Formation are biostratigraphically equivalent to upper or lower parts of the Eglwyseg Formation.

The presence of ? Lonsdaleia sp (Wedd et al. 1929, p.95) in the Llynolys Formation has not been corroborated here, but it was on the presence of these problematic specimens that they (op. cit) placed the upper part of the Llynolys Formation in the 'D_2'(Brigantian). From the presence of typical Asbian elements, and the lack of a distinctive Brigantian fauna, the Llynolys Formation is here considered to be wholly U. Asbian and within the Mesothem D5b (Ramsbottom, 1977).

4. STRATIGRAPHY AND CYCLE FORM OF THE TREFOR FORMATION.

Somerville (1977) proposed the term 'Trefor Limestone Formation' to encompass the 'Upper Grey Limestone' of Morton (1879) and considered it the lateral equivalent of the 'Cefn Mawr Limestone Formation' (Somerville, 1979C) of the Vale of Clwyd. Somerville did not, however, extend his minor cycle stratigraphic scheme into the Trefor (Limestone) Formation. He erected the type locality at Trefor Rocks (G. R. SJ 23054335 - 23204360),† defining its lower boundary with the underlying ' Eglwyseg(Limestone) Formation' (at the base of the prominent uppermost scarp feature above Trefor Rocks) at the first appearance of thin-bedded black argillaceous wackestones and shales, and its upper limit marked by a cross-bedded sandy oolitic grainstone (G. R. SJ 24054295), the basal unit of the Sandy Passage Beds (Somerville, 1977) (Morton's (1879) Sandy Limestone)). It is at its maximum of about 70m thick at the type section but thins southwards to ~25m in the vicinity of Whitehaven (Dolgoch , Plate 3,fig.B)

The formation is unique in this study in its continuous lateral extent from Minera, southwards to Llynclys, a distance of about 30km. The lateral persistence of facies across the Berwyn Hills made erection of an equivalent Oswestry area formation unnecessary. The uppermost beds of this formation are also exposed at the Hafod y Calch Inlier, Corwen, 12km west of Llangollen (G. R. SJ 053430). Chart D is a compilation of stratigraphic logs, abridged as a correlation guide in fig. 30.

Wedd et al (1929) believed the Upper Grey Limestone (equivalent to 'Trefor Formation') to be between 200 and 350 feet thick in the

---

† Plate 3 fig. A .

PULLOUT

Oswestry area and 300 to 800 feet thick between Llangollen and the Vale of Clwyd (although Morton (1879) published thickness figures similar to those presented in this work). The data presented here, based on a correlation scheme not previously developed shows that Wedd et al (1927) made a significant overestimate, probably induced by poor definition of outcrop pattern and the presence of repetitious faulting, especially in the Oswestry area.

In this stratigraphic scheme I propose a redefinition of the upper boundary of the Trefor Formation from evidence of minor cycle form and correlation.

4.1. STRATIGRAPHY OF THE TREFOR FORMATION

Stratigraphic horizons conform to the numbering scheme on fig. 30. Cycle boundaries are numbered with the prefix "Tf.".

4.1.1. Lower Cycles of the Trefor Formation.

The Trefor Formation is informally divided at boundary Tf.3 into a sequence of lower and upper cycles to facilitate their description.

4.1.1.1. SEDIMENTS BETWEEN BOUNDARIES Tf1 and Tf2. Above the uppermost Eglwyseg Formation cycle along Eglwyseg, a prominent scarp and talus slope represents the basal 10 to 20m of the Trefor Formation, thinning northwards from Trefor. The upper part of this is exposed at the southern end of Trefor Rocks (G. R. SJ. 230433), but is more fully exposed in the sequence of quarries extending southwards from Plas Ifa (G. R. SJ. 246424) (see fig. 30).

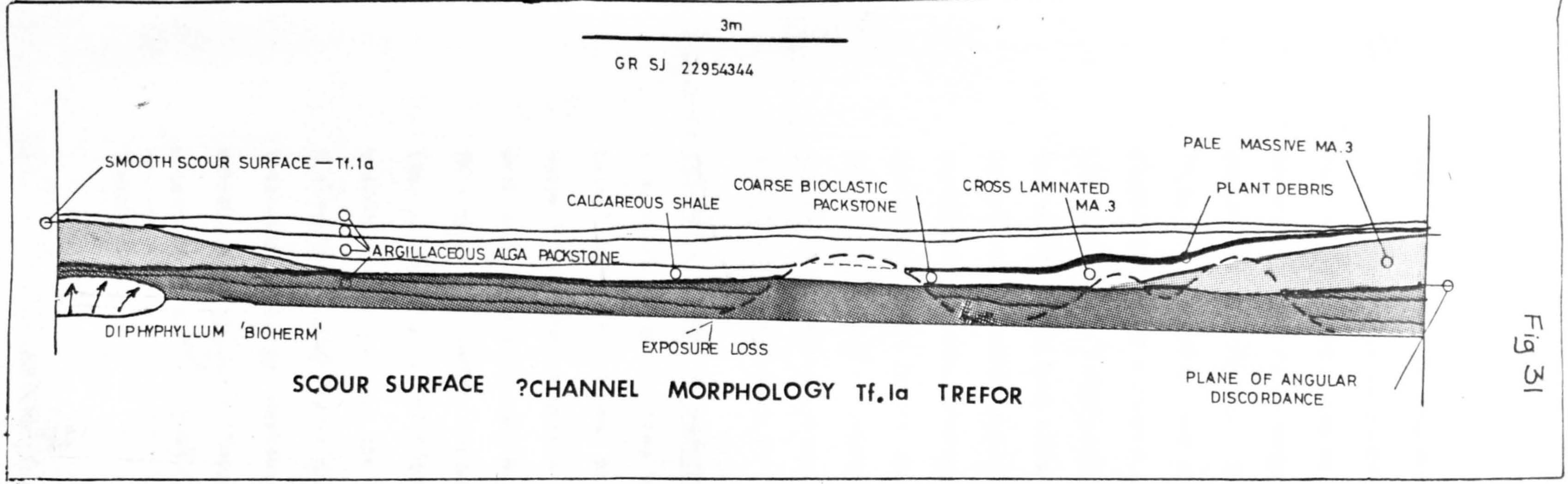
These sediments are black to dark grey, thin and wavy bedded (5 to 50cm) argillaceous packstone/wackestones, with interbedded shales and shale partings. They contain a diverse coral-brachiopod assemblage, including Diphyphyllum furcatum, Lithrostrotion irregulare, L. pauciradiale, L. maccoyanum, Lonsdaleia floriformis, Siphonophyllia sp., Aulophyllum fungites, Dibunophyllum bipartitum, Syringopora spp., Gigantoproductus spp., Productus (Dictyoclostus) spp., and Brachythyris sp.. The microfauna noteably includes Saccamminopsis sp., Tetrataxis conica, Valvulinella sp. and Loeblichia sp.. This assemblage is typically Brigantian.

Chert nodules are poorly developed, more often represented by thin silicified outer coatings of bioclasts, or as $\leqslant$ 10cm diameter 'honeycombe' nodules, where silicification has been only partial.

Small biohermal growths dominated by lithostrotiontids (see section 10. 4.) occur within the sequence at Trefor Rocks† (G. R. SJ23054335) and Pant Hir (G. R. SJ23772790). On Trefor Rocks, marking the upper boundary of this unit, is a small scale only locally recognised angular unconformity (Plate 3 fig. C), with angular discordance $\leqslant 5^{\circ}$. Underlying strata are slightly folded, although the unconformity is planar, without any apparent palaeokarstic relief or calcrete structures. Overlying this surface is a 0 to 60cm bed of medium grey bioclastic packstones, with a thin veneer of bioclastic laminated grainstone, capped by the 'smooth' surface Tf 1a. At G. R. SJ2294344 (Trefor Rocks) this bed has a local ?channelised primary relief, (fig. 31) and is absent southwards from Trefor. Whether this bed represents the upper phase of a minor cycle is unclear.

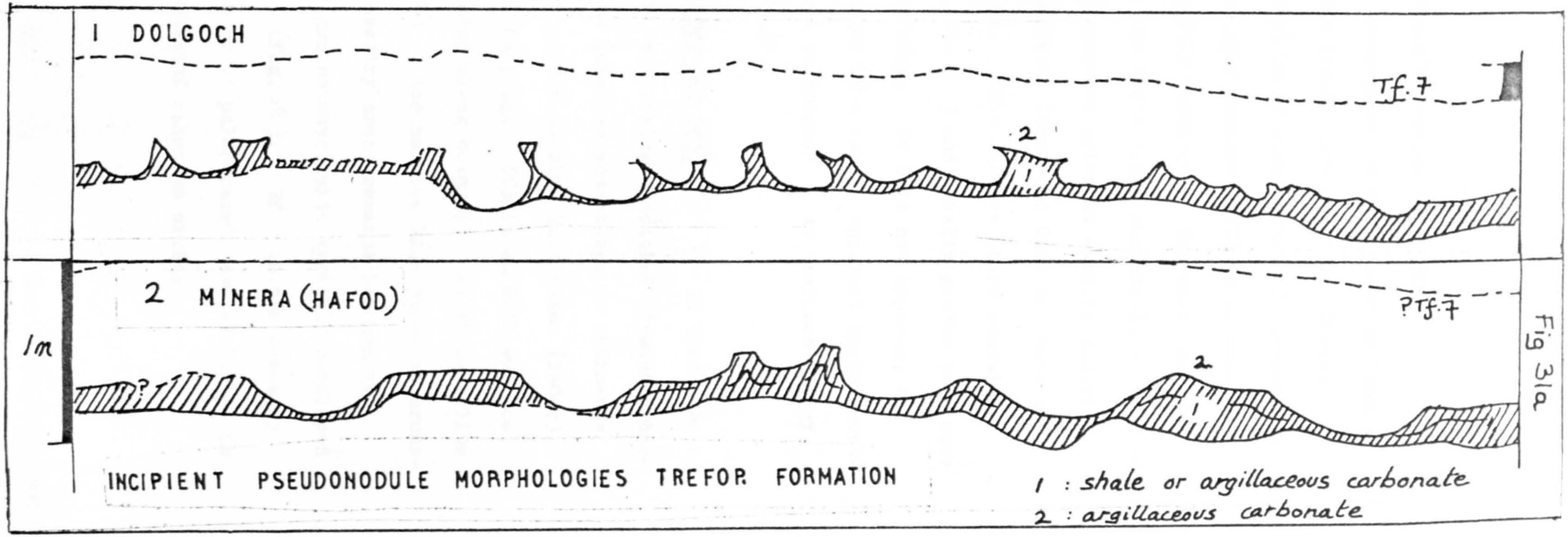
Tf.1a is overlain by 80 to 180cm of thin bedded packstones, containing a high diversity coral biostrome (section 10. 4.), including Palaeosmilia regia. Greater than 30% of the colonies are cerioid Lithostrotion maccoyanum. This biostrome can be traced over the 20km from Trefor to Dolgoch (fig.30). At Trefor Quarries the junction between this and the overlying massive phase

† Plate 18, Fig. G



SCOUR SURFACE ?CHANNEL MORPHOLOGY Tf.1a TREFOR

Fig 31



INCIPIENT PSEUDONODULE MORPHOLOGIES TREFOR FORMATION

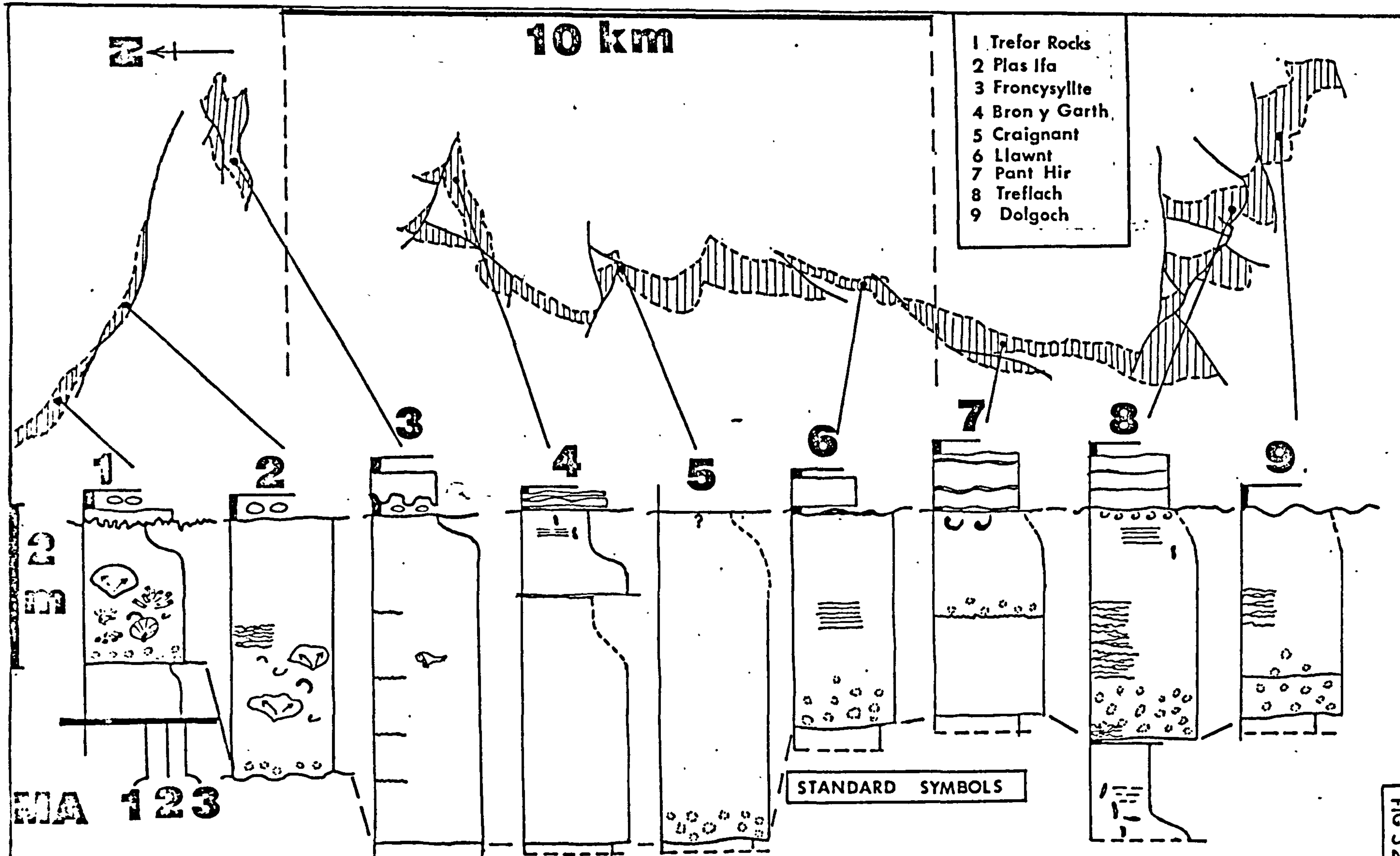
Fig 31a

is the site of 2cm deep vertical burrowing (s/m 220).

The lowest extensive development of MA.3 phase sediments, marking the upper phase of the first fully developed Trefor Formation minor cycle comprises ~1m of massive pale Coelosporella grainstone/packstones. The upper surface (Tf2) is similar to Tf1a, being, smooth, and slightly undulose. One coral colony planed off where embedded in the bed's upper surface (G. R. SJ23324329), indicates that lithification occurred prior to erosive scouring. The bedtop lacks true palaeokarstic form, and does not have any associated calcrete structures. This massive phase contains incipient bioherms (section 10. 4) and locally grades to a more micritic texture in its upper 10cm. Tf 2 is not apparent at Froncysyllte, and is represented by a rarely apparent cycle boundary of ?palaeokarst southwards to Dolgoch. At Treflach Quarry, a 2m MA.1 phase underlies 'Tf.2'.

4 .1.1.2. SEDIMENTS BETWEEN BOUNDARIES Tf2 and Tf3. At Trefor Rocks a 2m thick cycle overlies Tf2, comprising a mottled ('pseudobrecciated') lower biostromal phase (similar assemblage to sediments below Tf2) (20cm) overlain by a pale massive MA.3 phase (160cm), with a thin impersistent MA.1 capping. Tf3 is an S.D. surface†. This cycle is readily correlated along outcrop. At Froncysyllte (Pen y Graig, G. R. SJ26664061) the massive MA.3 phase is cross-bedded, but throughout the Oswestry area, remains homogenous Coelosporella packstones and grainstones, with variably developed laminated pelloidal horizons, (fig. 32). Tf 3 in the Oswestry embayment is either a 'Form A or D' palaeokarst, and at Llawnt is underlain by a thin (5cm) laminated calcrete crust.

† section 11.1



Regressive phase of Tf 2.3; Trefor Form'n

PLATE 3. Mesothem D6a.

A Type section of Trefor Formation, Trefor Rocks, G. R. SJ 233432. Minor cycle boundaries numbered.

B Dolgoch Quarry(Nut tree Bank) G. R. SJ 277247, Trefor Formation. Note thickness of each minor cycle c.f. fig. A. Note irregular 'loaded' nature of bed base within Tf. 6. 7.

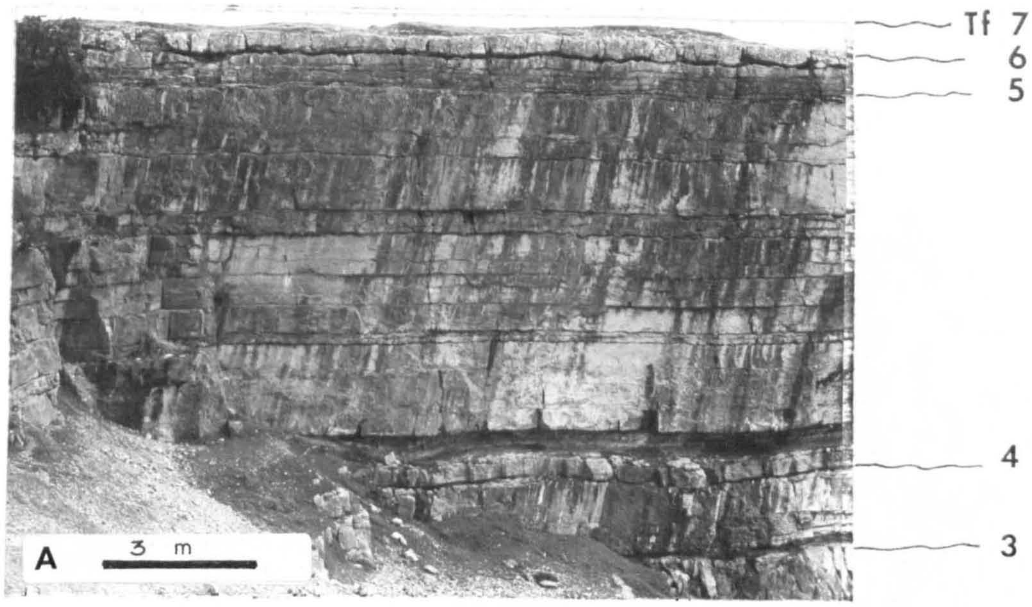
C Minor angular unconformity within basal transgressive phase of Trefor Formation, on Trefor Rocks. G. R. SJ 23054335.

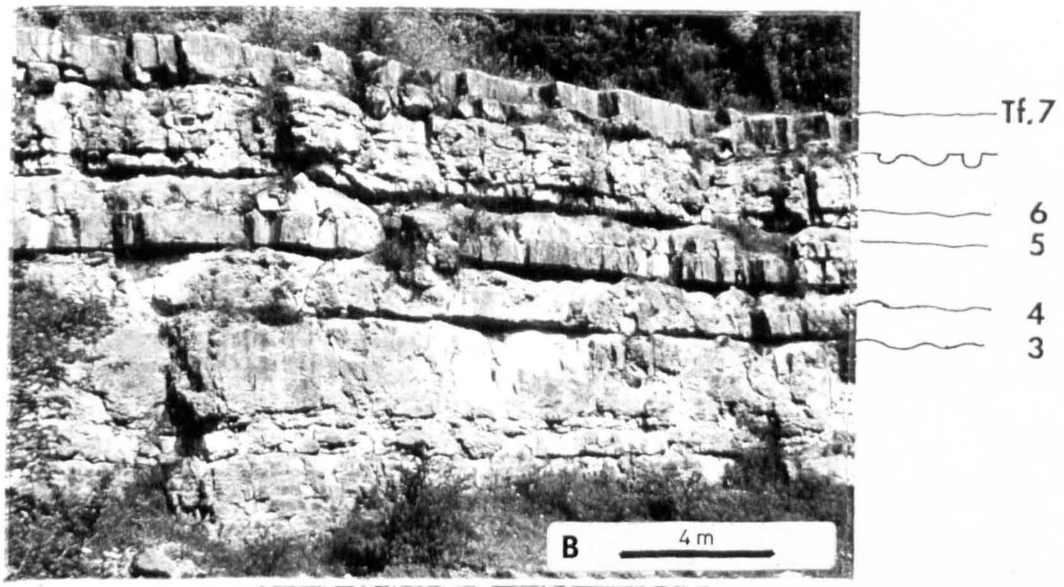
D Gigantoproductus sp biostrome at Minera, G. R. SJ 25275181

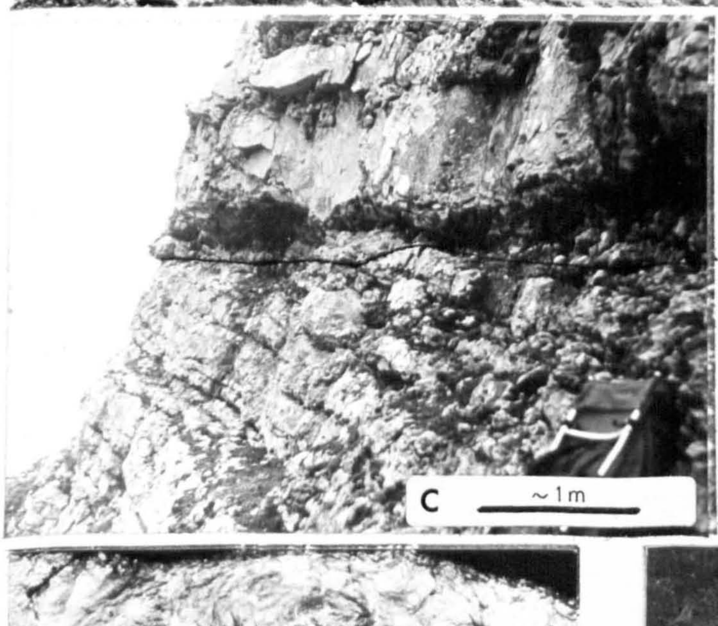
E Incipient pseudonodule structure and injection 'diapir', Tf.4.5., Trefor Rocks type section G.R.SJ23384328

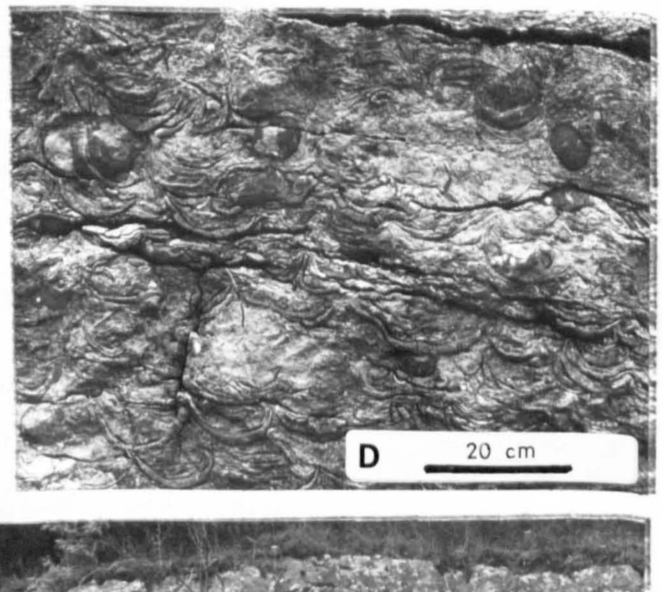
F Tf. 6. 7., Minera, with well exposed 'loaded' horizon at same stratigraphic level as loaded horizon of Plate2, fig. B. Tf. 6 is palaeokarstic, with underlying tubular calcrete structures and a thin laminated crust in G. R. SJ 25955174

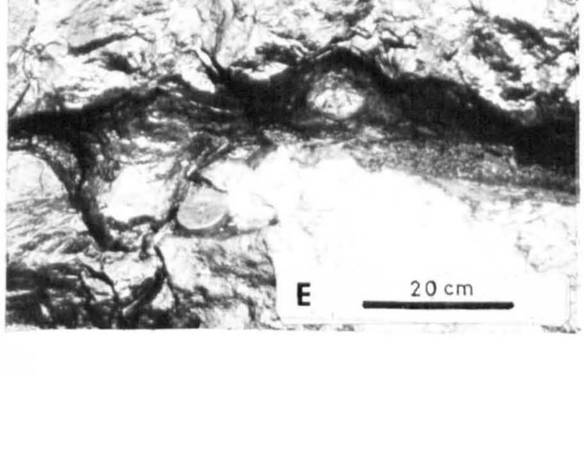
PLATE 3

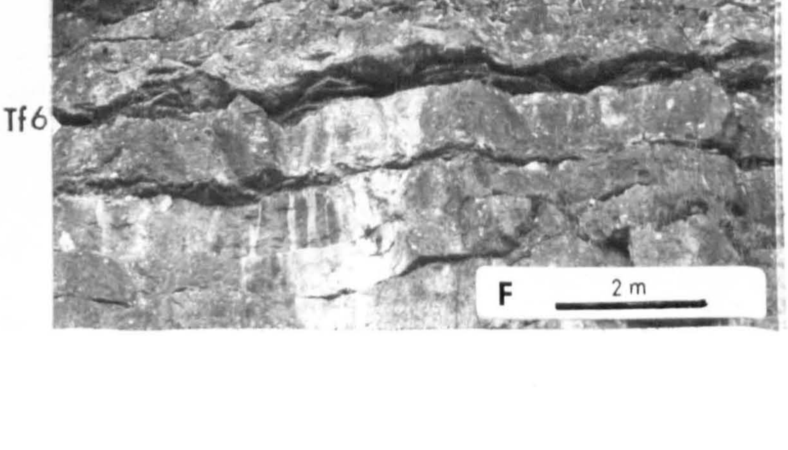












4.1.1.3. DEPOSITIONAL HISTORY OF LOWER CYCLES. The thickness of the sediment 'packet' below Tf 2 is variable in response to the underlying Caledonide tectonic elements, thinning over both the Berwyn and Cyrn y Brain axes. This is a continuation of the same trends in underlying Asbian strata. This thinning occurs partly by onlap of successive beds, and overstep of Asbian (and L. Palaeozoic) sediments.

The presence of an early Brigantian unconformity indicates that local tectonic influences were active. Walkden (1977) described an ?angular unconformity in early Brigantian strata from the Derbyshire Block. Further evidence of widespread earth movements during the latter part of the Asbian and early Brigantian comes from the Craven Basin. Hudson and Mitchell (1937) postulated three unconformities (B_2-P_{1a}; P_{1c}-P_{2a}; E_{1a}-E_{1b}) within the Bowland Shale Group. The former of these three is of a comparable age to the Tf 1 horizon (early Brigantian).

Sediments overlying the angular unconformity at Trefor Rocks, traced southwards, show little evidence of thickness variations over tectonic features and reflect a return to the shoaling cycle forms of the Asbian style. A change of available components (especially algal) and increased quantities of argillaceous materials (or, less likely reduced sedimentation rates) modifies the sediment microfacies with respect to their Asbian equivalents.

4.1.2. Upper cycles of the Trefor Formation.

4.1.2.1. SEDIMENTS BETWEEN 'Tf 3' AND 'Tf 4' At Trefor Rocks, and southwards these sediments comprise one whole cycle, varying from 2½m to zero. It appears† to be absent at Froncysyllte Pant Hir and Treflach. Towards the base a 50cm shale horizon is developed at

†(according to the correlation presented herein.)

Trefor Rocks, with thin nodular microsparite bands, succeeded by a thin-bedded MA.2 phase and a massive MA.3 bioclastic grainstone phase. An impersistent MA.1 phase caps the cycle at Trefor, but is absent elsewhere.

The surface 'Tf4'undulates irregularly with a low amplitude ($\leq$15cm), but overlying wavy-bedded argillaceous wackestone/packstones have similar relief. Its association with MA.1 and MA.3 microfacies is the only readily distinguishing ?palaeokarstic criterion.

Northwards from Trefor correlation of this cycle is tentative. It may be laterally equivalent to 1m of laterally impersistent crinoidal rudstone at Minera.

4 .1.2.2. <u>SEDIMENTS BETWEEN 'Tf 4' AND 'Tf 5'</u>. Representing one cycle, this sediment packet is thickest along Trefor Rocks (11m), and thins southwards to 2m at Dolgoch. A shale bed, the thickest exposed at Trefor Rocks (1m) is persistent across outcrop, towards the cycle base, containing microspar nodules, many enveloping uncrushed fasciculate lithostrotiontid colonies that indicate nodule formation was prior to clay compaction. This is succeeded by a sequence of thin wavy bedded (5 to 20cm) and more massive bedded argillaceous wackestones that thin from 8m at Trefor Rocks to zero southwards. Many of the bedding planes in this unit have a bizarre morphology, probably as a result of pre-lithification loading and dewatering (see on), producing incipient pseudonodules (Plate 3 fig. E). The unit contains laterally impersistent fasciculate listhostrotiontid dominated biostromes (especially <u>L. junceum</u>, and locally supports <u>Gigantoproductus</u>/<u>Gigantella</u>/<u>Semiplanus</u> shellbeds (especially well developed at Minera (Plate 3 fig. D)).

The uppermost MA.3 dominated phase of this cycle is variable from homogeneous Coelosporella grainstone-packstones to laminated and mottled Coelosporella-peloid wackestone-grainstones. A Gigantella shell bed characterises its upper surface between Pant Hir and Dolgoch. 'Tf5' is a smooth to undulose 'palaeokarst form D', lacking calcrete phenomena. (See Chapter 11.)

4.1.2.3. SEDIMENTS BETWEEN 'Tf 5' AND 'Tf 6' This cycle is dominated by a thick upper MA.3 and MA.1 unit. At Trefor and Garth Obry (G. R. SJ 258380) the cycle base is marked by a thin 100to 30cm MA.2 phase, grading through a mottled zone into massive bedded MA.3 sediments. This basal phase, between Llawnt and Pant Hir has a fasciculate lithostrotiontid dominated biostrome (L. irregulare, L. pauciradiale, L. junceum, L. cf maccoyanum (fig. 30.)) whilst within the upper MA.1 dominated unit at Trefor Semiplanus dominated brachiopod assemblages and a Chaetetes / Syringopora association are present. Within this MA.1 phase, at both Garth Obry and Trefor a prominent palaeokarst ($\leq$ 50cm relief, Form A†) is developed with an impersistently preserved laminated crust. As with many 'cycle' boundaries in the underlying Asbian, multiple boundaries reflect minor scale cyclic events. This cycle is apparently absent at Minera.

4.1.2.4. SEDIMENTS BETWEEN 'Tf 6' AND 'Tf 7' The base of this cycle comprises 20cm $\rightarrow$ 5m of interbedded shales and thin bedded MA.2, with a Gigantella-Semiplanus brachiopod assemblage at Dolgoch. A prominent 'convolute' surface occurs in this basal phase, traceable at the same shale/limestone lithological transition, along outcrop from Minera to Dolgoch (30km). Large incipient (not totally detatched) pseudonodules up to 50cm diameter‡ (Plate 3 fig. B&F) penetrate the underlying shale. No lamination structure is visible in the lime-

---

† Chapter. 11 ‡ Fig. 31a

stone, although shale laminae do only, in part, circumscribe the 'nodules', more through compaction, rather than original disruption(?).

The thin (? ⩽ 1m) MA.3 phase, overlying this load-bedded unit is only exposed at Trefor Rocks, Llawnt and Dolgoch, and is capped by a variable ⩽ 2m MA.1 development that form the top of many natural exposures from Minera to Dolgoch. At Trefor Rocks the regressive phase of this cycle records two emergence events, separated by 80cm of MA.1.5. These two surfaces presumably converge southwards. At Dolgoch 'Tf.7' is a palaeokarst of 'Form D' without any calcrete structures associated.

SEDIMENTS BETWEEN Tf.7 & Tf.8 In the Oswestry area the base of this minor cycle is a conspicuous three- five metre shale, correlated by Morton (1879) as approximately equivalent to the Coral Bed at the top of the Upper Grey Limestone at Llangollen.

The presence of an overlying MA.3 and MA.1 unit at Llawnt, Craignant, and Bron y Garth, and not the sandy carbonate facies above Morton's Coral Bed at Trefor, suggests this correlation of Morton's is invalid.

Beneath Tf.8 at Trefor Rocks is a well-developed 2½m MA.1 unit. This makes a low scarp traceable northwards to Llwyn Hen Parc.. An underlying veil may represent the thick shale level.

Tf.8 is only exposed at Trefor Rocks (G.R. SJ 2353 4360), there notably with a 2cm laminated crust immediately underlying.

4.1.2.5. SEDIMENTS ABOVE Tf.8 (SANDY PASSAGE FORM'N) Sediments above 'Tf. 8' are poorly exposed along outcrop. At Trefor and northwards a prominent scarp above 'Tf.8' marks the position of Morton's (1879, p.38) coral bed. The top 8m of this scarp are exposed in a series of 'quarries' at Eglwyseg Plantation (G. R. SJ 227440) and as minor outcrops along the scarp between Tynant and Llwyn Hên Parc Ravines , comprising thin-bedded argillaceous packstones (L. 3) (MA.2) and containing the most prolific and diverse coral-brachiopod fauna recovered from any one Trefor Formation cycle. Small coral biohermal developments also occur within (section 10. 4.). Significant elements include: Palaeosmilia regia; Lonsdaleia duplicata; Lonsdaleia floriformis floriformis (small form of Smith (1915); Clisiophyllum spp; Dibunophyllum bipartitum subspp. ; Corwenia rugosa; Lithostrotion edmondsi and Syringopora catenata. The full diversity is given in appendix IV. This unit correlates (?) with 'minor cycle 7' of Somerville (1979c) from the Cefn Mawr Limestone of the Mold area.

(Wedd et al. (1929, p. 99-100) recorded a Lonsdaleia duplicata coral fauna from sediments a?few metres above Tf.8 in the Oswestry area, suggesting that Morton's Coral Bed extends southwards a considerable distance. They (op. cit.) included this, however, within their Sandy Limestone.)

Overlying the MA.2 phase is a massive-bedded unit of coarse grainstones and rudstones, with coarse sand and pebble grade quartz lithoclasts. This was taken by Morton (1879) as the base of the 'Sandy Limestone'. Somerville (1979a, p.337) defined the top of the Trefor Limestone as marked by the appearance of a cross-bedded sandy oosparite unit developed above the coral bed. At G. R. SJ 23034373,2-4m of cross-laminated and massive oolitic and quartzose grainstones underlie the cross-bedded unit of Somerville. This 2-4m of arenaceous carbonates is here included within the Sandy Passage Formation.

It is at this equivalent level across the study area (? and at Corwen, 12km west) that the arenaceous developments commence. Significantly, these carbonate-clastic units represent the regressive phases of minor cycles. Therefore the Trefor Formation boundary, as depicted by Somerville lies within a minor cycle. Unfortunately, exposure is too poor along outcrop to elucidate the Sandy Passage Formation minor cycle stratigraphy, although the scattered exposure suggests that carbonate-clastics are characteristic throughout. This cycle form is more comparable to the Yoredale cyclic clastic-carbonate facies of North Yorkshire, both in age (Upper Brigantian)?and form.

To place significance on the Trefor Formation/ Sandy Passage Formation boundary, I propose that it is placed at a minor cycle boundary, between cycles of different character. The Trefor Formation upper boundary is therefore placed along Tf.8 in this study†, conforming to the Upper Grey Limestone / Sandy Passage Bed boundary of Wedd *et al*(1929).

4.1.3. Penecontemporary Tectonic Soft Sediment Deformation.

The lateral persistence of loaded surface morphologies is significant. Most other shale/limestone contacts are either planar, wavy (conforming to the bed morphology of the thin bedded MA.2 phase), or only slightly loaded (relief ≤ 15cm)(cf fig 31a).

Coulter and Migliaccio (1966) noted that 'soils' suffer loss of bearing capacity when subject to seismic shock. The result is an increase in pore water pressure, and liquifaction.

Sims (1975) described deformational structures in lake silts resulting from earthquake shocks. The principle structure observed in his study was an incipient pseudonodule. He (*op.cit.*,p.146-147) correlated five zones of deformed strata over 100km^2. These structures are, however, small scale, and best developed in matrix-free sands and silts.

The dark argillaceous packstones and wackestones of the Trefor Formation do not present as readily a fluidised sediment, although Lowe (1975, p.175) noted that "unlithified clays and muds and cohesionless sediments rendered quick by rapid seepage, liqui-

† Type section redefined as G.R.SJ 23054335 - 23534360

faction or fluidisation, flow readily if sheared... [producing]... soft-sediment intrusions".

There is no evidence of a palaeoslope in the Trefor Formation succession along the Minera-Dolgoch outcrop that may have induced soft sediment flowage. The 'load' morphologies and distribution do not suggest horizontal movement (e.g. there are no observed recumbent folds cf. Allen and Banks, 1972).

The lateral extent of these Brigantian 'loaded' horizons and the implication that tectonic movements were occurring due to both the overall tectonically influenced sedimentation pattern, and presence of internal unconformities suggests that fluidisation was induced by seismic activity.

4.1.4. Depositional History of the upper cycles of the Trefor Formation.

The thinning of strata over structurally ' positive' 'basements' as occurred in sediments below Tf 1a is less noticeable in higher strata. Most prominent is a general thinning southwards from the Trefor area (central to the Llangollen embayment) to Dolgoch. This thinning is not however, primarily associated with cycle offlap, but by thinning of individual cycles, and primarily the reduction in thickness of basal MA.0 and MA.2 phases. Cycle 'Tf. 3/4 ' is exceptional in being unrepresented at certain localities.

The uniformity of cycle's upper phases across the region is striking (e.g. fig. 32) although this is in part controlled by the overall depositional setting of the outcrop pattern being parallel to depositional strike.

The thick unit of thin bedded argillaceous sediments above 'Tf. 8' placed here as the basal unit of the Sandy Passage Formation, may represent the first cycle within the Mesothem D6b. of Ramsbottom (1977), due to the thick nature of its transgressive phase with respect to immediately underlying minor cycles.

4 .2. CYCLE FORM. (Refer to Fig. 33)

Cycles of the Trefor Formation have similarities with both the Tynant / Sychtyn cycles and those of the Eglwyseg Formations. They represent a shoaling sequence of facies. In most cycles basal thin-bedded MA.2 (Lithofacies L.3) dominate, accounting for about 70% of the measured section at Trefor. Lithofacies L. 3 fall into two bedtypes: those wavy and thin bedded separated by thin shale partings, and those thin-bedded but separated by small irregular stylolites. These two bed-types are interstratified, and in cycle 'Tf 4.5' (e.g. Hafod, Minera and Trefor, see chart D) they are 'cyclically' repeated as bed-sets.

The most extensive and fossiliferous coral-brachiopod biostromes occur within these sediments.

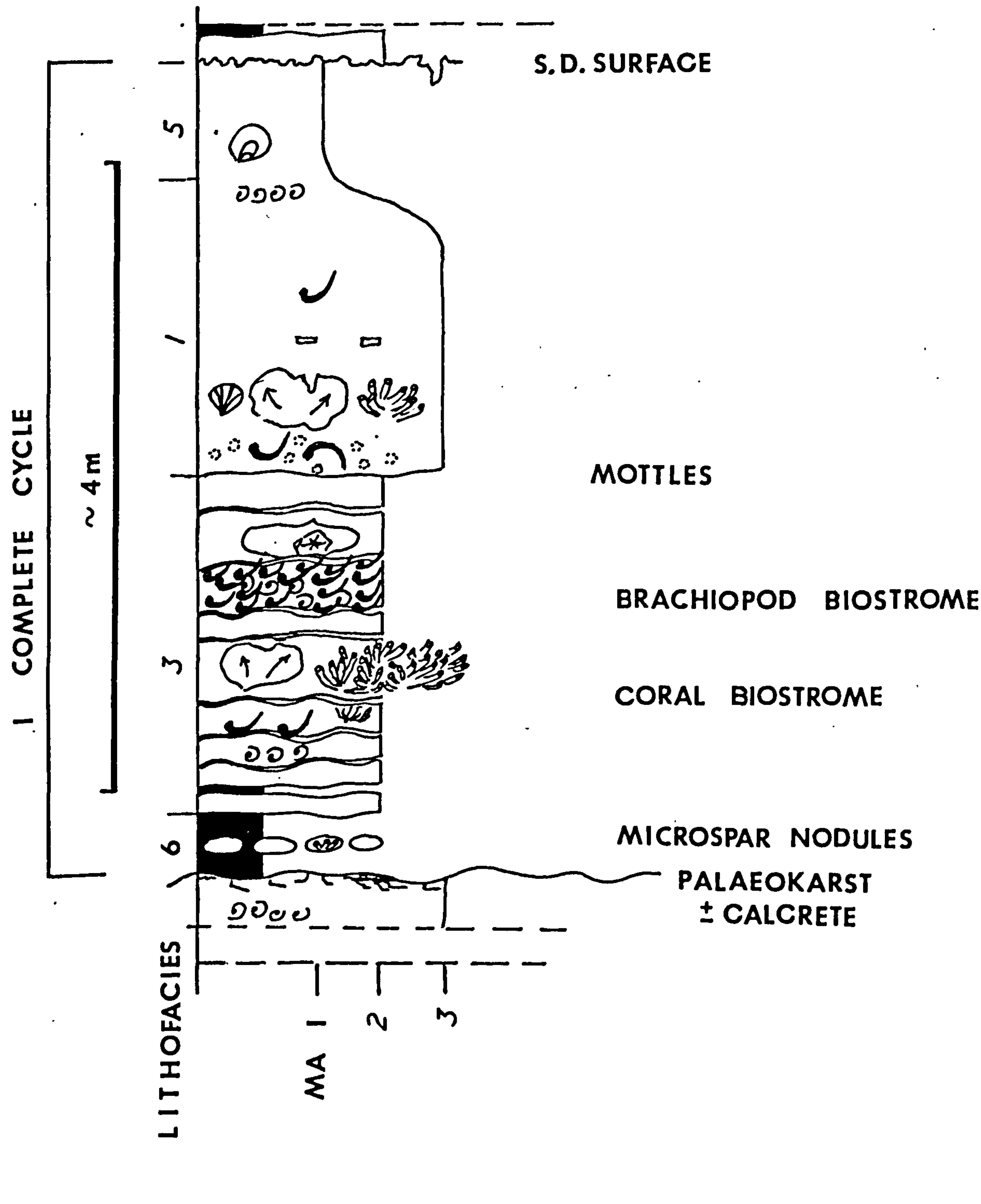
Paler, massive bedded MA.3 phases account for about 10% of the Trefor succession, but greater than 20% at Dolgoch. They often have 'pseudobreccia' mottled bases and are <u>Coelosporella</u> and peloidal grainstones and packstones. With increasing silt and mud grade matrix this phase grades into a variably developed MA.1 phase, of characteristic calcisphere wackestones.

These MA.1 sediments lack the petrographic variation of their Asbian counterparts, and few have been observed with fenestral fabrics, early cements or penecontemporary dolomites, although S.D. surfaces attest to their early lithification and both calcrete phenomena and palaeokarsts of 'Forms A and D' attest to their subaerial emergence, (see chapters 10 and 11).

4 .3. PALAEONTOLOGY.

The presence of a Lonsdaleoid coral fauna, along with such forms as <u>Lithostrotion decipiens</u> , L. <u>pauciradiale</u>,

COMPOSITE SEQUENCE TREFOR FORMATION
MINOR CYCLE



Palaeosmilia regia; Aulophyllum fungites and Lonsdaleia floriformis is diagnostic of the Brigantian. The

The lowermost minor cycle of the Sandy Passage Formation, conformable with underlying Trefor Formation, contains a diverse and characteristic mid Brigantian assemblage, including such forms as Orionastraea tuberosa; Corwenia rugosa and Nemistium edmondsi. (see Appendix V).

5. COMPARISON WITH ASBIAN OF THE ASKRIGG AND RAVENSTONEDALE AREAS

The northern margin of the Craven Basin (fig. 3) is defined by the Craven Fault Belt. North of this active tectonic hinge, during the Lower Carboniferous, the positive 'Askrigg Block' periodically provided a shallow shelf environment for carbonate deposition.

It is the presence of late Asbian deposits ascribable to Mesothem D5b (Kingsdale Limestone) that makes this area important for comparison of palaeoenvironments with North Wales, especially as it may be considered a tectonic unit isolated from St. George's Land. The Kingsdale Limestone sequence, therefore, provides an ideal comparative for study of the minor cycle character on the two shelf areas, in order to assess the eustatic component in minor cycle formation.

To the North of the Askrigg Block, the east - west trending Ravenstonedale - Stainmore 'Trough' (fig. 3) was transgressed by early Asbian seas of Mesothem D5a, which deposited the Potts Beck Limestone. This area is also palaeogeographically and tectonically separated from North Wales and, therefore, is an ideal comparative early Asbian sequence.

5.1. THE POTTS BECK LIMESTONE (Mesothem D5a).

This formation was erected by George _et al_ (1976) to encompass early Asbian sediments of the Ravenstonedale / Stainmore trough, underlying the Upper Asbian Knipe Scar Limestone, and overlying the Holkerian Ashfell Limestone.

This Formation is best exposed beneath the Upper Asbian Knipe Scar Limestone in the valley of the River Cleugh, G.R SD 694916 to SD 696913 . Limited to the confines of the Ravenstonedale / Stainmore Trough, and Mesothem D5a, it is supposedly represented by a major non-sequence (Ramsbottom (1974) above the Porcellanous Limestone (and top of the Holkerian) of the Askrigg Block and by a non-sequence northwards along the western margin of the Alston and Cumbria successions (George et al 1976).

5.1.1. Lithostratigraphy

The Potts Beck Limestone comprises a thick sequence (113m logged, but this may include some Holkerian Ashfell Limestone towards the base) of argillaceous and thin bedded biosparites and biomicrites,(Lithofacies L.2 and L.3) alternating with porcellanous micrites (of Lithofacies L.5). interbedded in a cyclic manner similar to the Tynant Formation (fig. 10). Within the middle of the logged succession is a 38m thick development of Lithofacies L.3. Lenticular cherts are common throughout this sequence.

MA.3 sediments tend to be lighter in colour than the dark grey or black argillaceous MA.2 and MA.1.

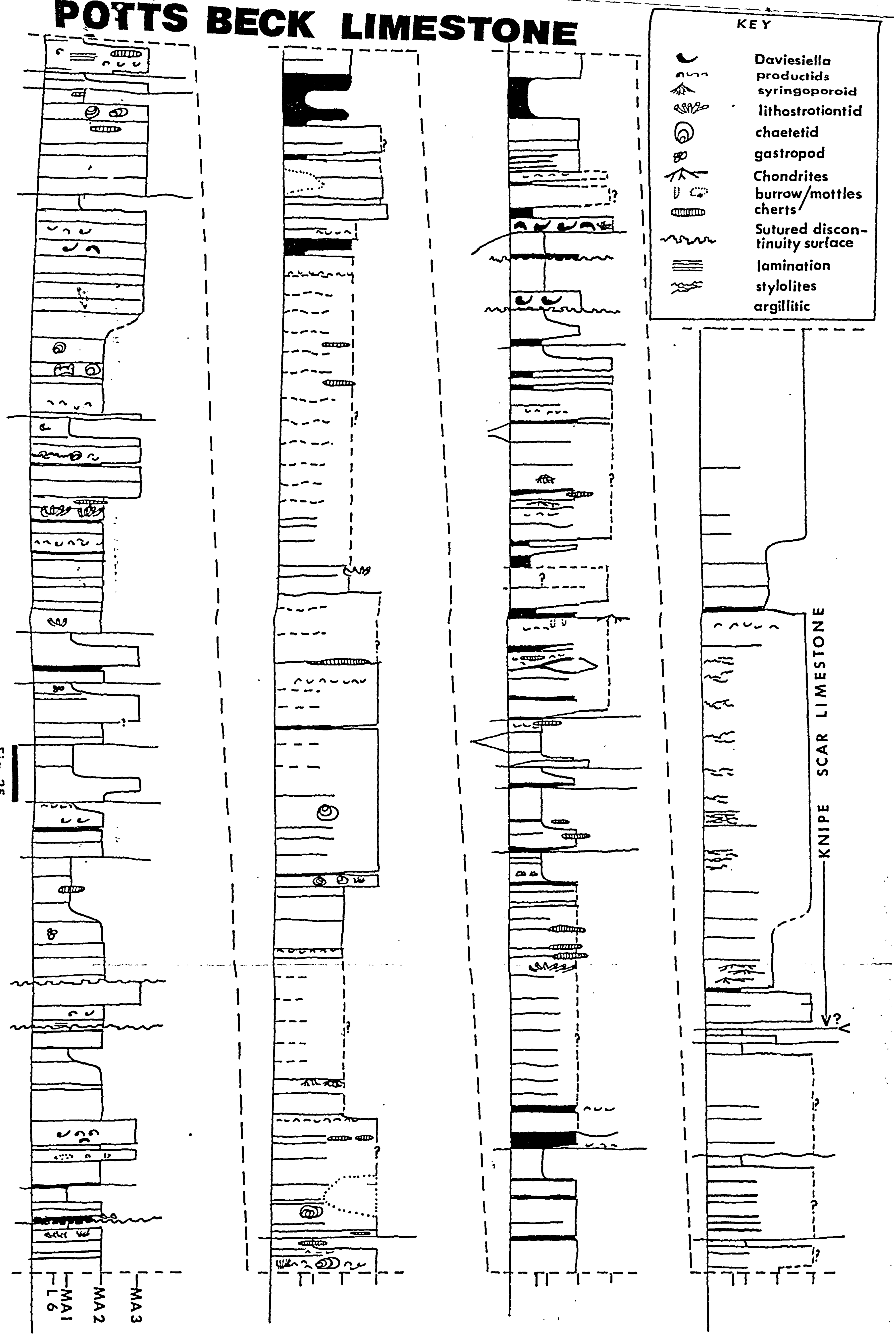
Cycle tops, marking regression / transgression boundaries are either planar surfaces or 'sutured discontinuity surfaces' (sensu chapter 11) ("burrowed hardgrounds" of Ramsbottom, 1974, p.59).

At least 20 minor cycles (individually > 1m thick) have been recognised within the logged section (fig. 34).

The Knipe Scar / Potts Beck Limestone transition was not apparently a non-sequence but comprises 15m of Lithofacies L.2 with 4 thin (≤ 50cm) beds of porcellanous micrites, underlain by 4 minor cycles with prominent regressive phases. This transition is rem-

FIG 34

POTTS BECK LIMESTONE



RIVER CLEUGH GRID REF. SD 694916 – 696913 inisoent of the Tynant / Eglwyseg Formation boundary in the vicinity of Tynant, suggesting the upper boundary of the Mesothem D5a, and top of the Potts Beck Limestone is at the top of the 4 minor cycle regressive phase (fig. 34). Noteably Daviesiella llangollensis occurs immediately overlying this horizon, as in the basal Eg.1.2 Eglwyseg Formation minor cycle at Llangollen.

5.1.2. Cycle Form.

Most of the cycles are 3 to 10m thick, with basal argillaceous and thin bedded Lithofacies L.3 (MA.2) overlain by variably developed 20 - 100cm bedded Lithofacies L.2 (MA.3), with thin porcellanous micrites (L.5) (MA.1) upper phases. Fig. 35 shows a detailed component log across a cycle from the lower part of the sequence (position marked on fig. 34). Compositionally, the sediments follow parallel trends observed in the Tynant Formation (section 2.1.1) and plot in a similar position on the C-A-M tenary diagram (section 2. 1. 2 ; fig. 36), to the Tynant Formation sediments.

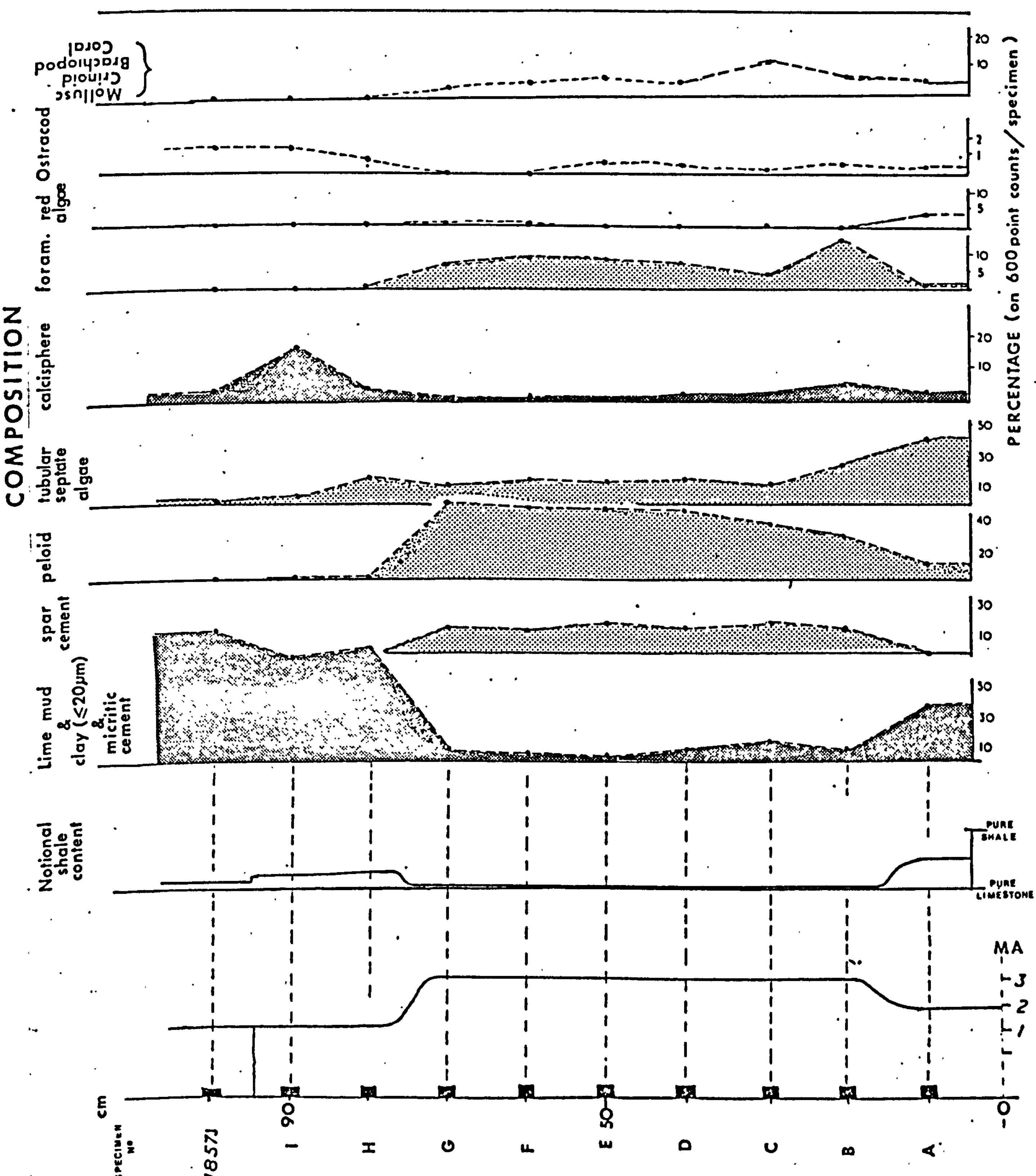
No palaeokarsts occur within the succession. Cycle transitions are either planar (?erosion) surfaces or sutured discontinuity surfaces (section 11.1.).

5.1.3. Significance of the Potts Beck Limestone

The River Cleugh succession of the Potts Beck Limestone, in showing similarity to the Tynant Formation, indicates that this cycle form has a more cosmopolitan uniformity.

The Ravenstonedale 'Trough' was apparently subsiding relative to the surrounding positive regions of Alston, Askrigg and the Lake District.

Detailed log of a minor cycle in the basal Potts Beck Limestone ; River Clough , Sedbergh.
SD 695914



TERNARY PLOT OF
BIOCLAST COMPOSITIONS
POTTS BECK LIMESTONE

tie lines indicate
superposed samples

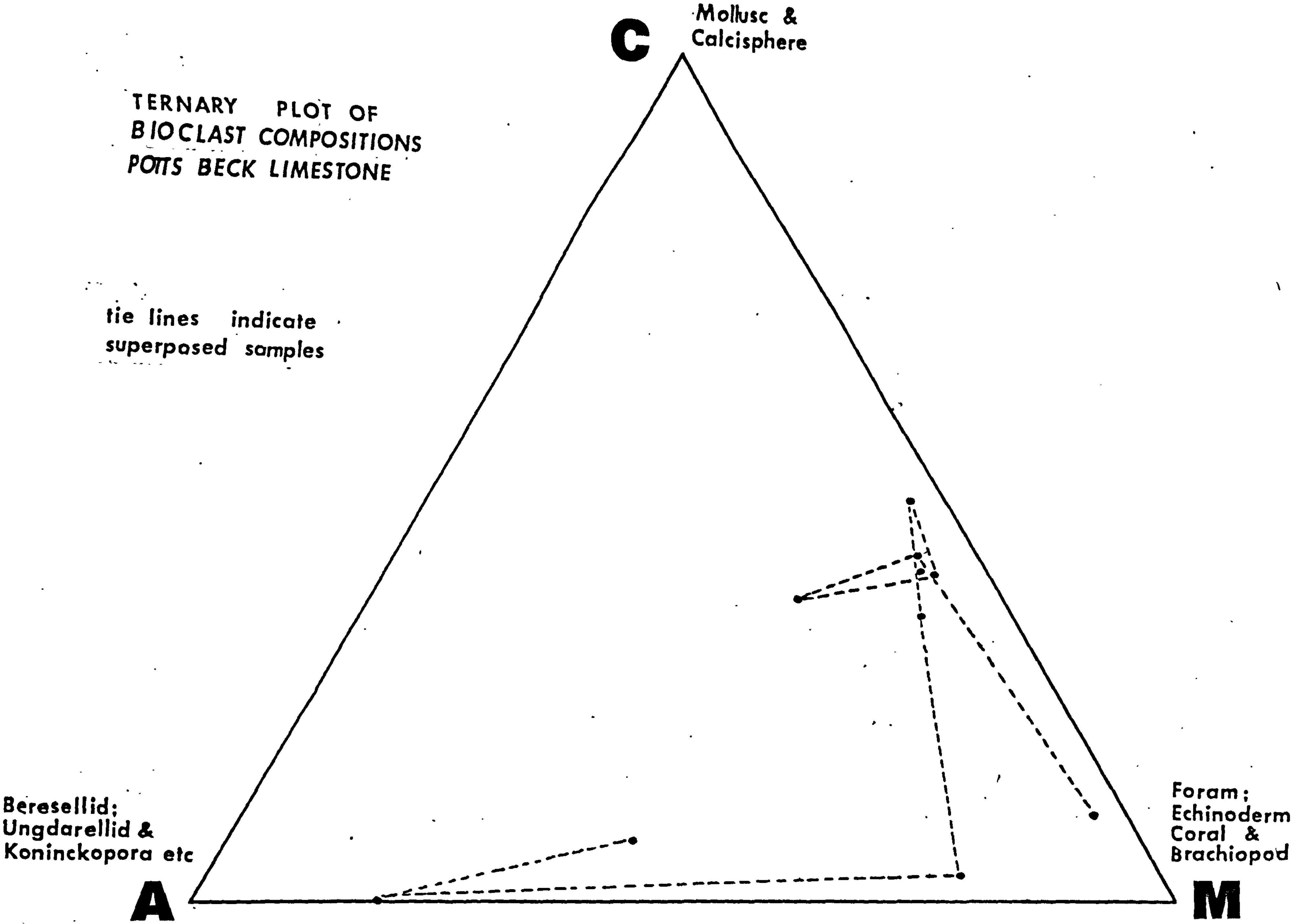


FIG 30

The lack of palaeokarsts, and rareity of both fenestral fabrics and sutured discontinuity surfaces suggests that, like the Tynant Formation, periods of subaerial emergence were limited in duration (and extent). Many of the regression / transgression boundaries are planar surfaces and may be subtidal in origin.

The lack of subaerial emergence, and the major development of subtidal (and transgressive) phase sediments within the mid part of the sequence suggests that subsidence played a more dominant role in the sedimentational history of the Ravenstonedale area than the Llangollen region.

5.2. THE KINGSDALE LIMESTONE (Mesothem D5b)

The Kingsdale Limestone represents the upper Asbian on the Askrigg Block. Northwards, strata equivalent to upper beds of the Kingsdale Limestone are differentiated as Yoredale 'cyclothems.'

Its base is marked by the 'Porcellanous Bed' - an impersistent regressive phase calcisphere wackestone - marking the top of the Holkerian. The early Asbian Potts Beck Limestone of Ravenstonedale is apparently absent on the Askrigg Block.

Exposed upon Twistleton Scars, the "Porcellanous Bed" is a multiple unit, with at least 3 thin 'cycles' represented by regressive phase calcisphere wackestones. The underlying Horton Limestone rests with angular unconformity upon the basement Ingletonian.

On Twistleton Scars, the Kingsdale Limestone is about 75m thick (104m beneath Wernside according to Waltham, 1971) but in Meal Bank Quarry, Ingleton (G.R. SD 699737) 4km south, and between the Middle and North Craven Faults, the lower boundary of the Kingsdale Limestone is uncertain due to the absence of the

Porcellanous Bed, although Garwood and Goodyear (1924) estimated its equivalent chronostratigraphic position 20m above the base of the Meal Bank succession (from the presence of Lithostrotion minus).

The Kingsdale Limestone comprises cycles (first recognised by Schwarzacher, 1958 by bedding plane distributions) of carbonate deposition and subaerial emergence, with associated calcretisation and karstification. The Meal Bank succession comprises at least seven such emergence horizons, including the lowest, well documented palaeokarst with infilling clay and coal.

The emergence events are analogous to those of the Eglwyseg Formation (of equivalent age, at least in part). The deeper palaeokarst reliefs lack calcrete phenomena, whilst most karsts (of form 'A', section 11.2.1.1) have tubular calcrete structures underlying. Only one surface was observed with a laminated crust poorly preserved on high points of the surface (Chart B).

The lithological variability of these cycles, unlike many of the Upper Asbian of Llangollen, is less marked.

Basal phases of these Kingsdale Limestone cycles are either 'pseudobreccia' mottled massive or stylolite-bedded Lithofacies L.1. or stylonodular and argillaceous Lithofacies L.4. The thin bedded Lithofacies L.2 and L.3 transgressive phases of Eglwyseg Formation cycles are lacking. The 'cyclicity' retains the more homogeneous character of the Eglwyseg Formation as developed on the positive 'axis' of the Cyrn y Brain.

Implications.

The Askrigg Block was (and to a limited extent, still is) a positive area during Lower Carboniferous deposition, with respect to the surrounding 'basin' or 'trough' situations. The Ingleton succession is situated on the tectonically active (?) southern margin of this block, associated with the Craven Fault system.

The lack of basal argillaceous transgressive phases to the Kingsdale Limestone minor cycles suggest that the earliest effects of the transgressive seas were not apparent on the shelf, or that subsidence effects were not marked enough to cause the transgression to exceed the depth to 'fair-weather' wave base.(see p.140).

The Askrigg succession was initially investigated to see whether there was any possibility of cycle correlation across the Craven Basin to other 'positive' regions during the Asbian.

The results obtained are:

1) On the S. margin of the Askrigg Block, near Ingleton, fewer major emergence events occurred in the U. Asbian cf. Llangollen, but this emergence resulted in the operation of similar palaeokarstic and calcretisation processes.

2) No direct cycle for cycle correlation can be at present erected between North Wales and Ingleton due to the lack of specific definition of each or any particular cycle.

Part C

MICROFACIES ANALYSIS AND PALAEOENVIRONMENT INTERPRETATION

6. PERITIDAL DEPOSITS IN THE TYNANT FORMATION AND SYCHTYN MEMBER

Microfacies represented within the peritidal sediment suite of the Tynant Formation and Sychtyn Member:

Calcisphere wackestone microfacies association (MA.1):

- MA.1.1. Calcisphere packstone to mudstone.
- MA.1.2a Unlaminated calcisphere-peloid grainstone/packstone.
- MA.1.2b Laminated calcisphere-peloid grainstone-packstone.
- MA.1.3. Cryptalgal laminite.
- MA.1.4. Laminated beresellid calcisphere mudstone packstone.

(Intraformational conglomerate lithofacies (L. 7)

- L.7.1. Clast supported conglomerates.)

6.1. Distribution and Field Character of MA.1

Calcisphere wackestones (MA.1) occur throughout the Tynant Formation and Sychtyn Member. They are normally present as the dominant regressive microfacies association towards the upper boundaries of cycles (see section 2. 1. 1) and are most common in these two units, locally contributing up to 50% of the succession (see correlation charts A & C). They also occur to a lesser extent in all the other formations and members recognised herein.

MA.1. bed morphologies vary between small-scale laterally impersistent units to larger-scale units, (see fig. 37) over 10m thick (e.g. top cycle of Tynant Limestone) that are readily correlated across kilometres of outcrop.

Emergence features (palaeokarsts, palaeosols and sutured discontinuity surfaces†) are often associated with their upper surface, whilst they usually grade downwards into sediments of MA.3 or MA.2 suite.

Sediments of MA.1 in hand specimen vary from dark grey or grey-

† see Chapter 11.

blue to pale cream-grey, and break with a conchoidal fracture. They correspond to the "porcellanous micrites" of Lithofacies L. 5.

Low diversity mollusoan-dominated macrofossil assemblages are characteristic, fenestral fabrics very frequent and penecontemporaneous dolomites restricted to this microfacies association.

6.2. PETROGRAPHY OF CALCISPHERE WACKESTONE (MA.1) AND ASSOCIATED SEDIMENTS

6.2.1. Petrography of MA.1.1. Calcisphere Packstone to Mudstone (Plate 4 fig. A)

This is the dominant lithology within MA.1. Wackestone is the most common textural type. The sediments are usually homogeneous, with rarely developed vague millimetre to centimetre lamination, and vague centimetre colour mottling with diffuse mottle margins. The dominant bioclasts are calcispheres and molluscs (gastropods). Small and medium irregular fenestrae are common, but tubular and 'packing' (new term, see section 6. 4. 1.4) varieties also occur. Petrographically this microfacies grades into, and may be interlaminated with other MA.1. microfacies, MA.2.1 or MA.3.1 (see Chapter 7). Variable volumes of insolubles are present dominated by illite (Appendix $\overline{\text{III}}$) although locally, diagenetic quartz prisms (c. 250 µm) are abundant e.g. s/m 388.

The matrix is dominantly lime-mud and silt-grade bioclastic material. In very thin section the 'lime-mud resolves to a clotted texture due to the presence of many ~30µm diameter peloids floating or packed in 3µm to 5µm micrite (Plate 4 fig. C), also noted by Schwartzacher (1961). These peloids are of two forms: those containing bioclastic silts and micrite grains, with diffuse margins (Plate 4, fig. C.) and probably of faecal origin; irregular, more dense micritic peloids with sharp margins probably derived from degraded bioclasts (Plate 4, fig. D). Hardie and Ginsburg

PLATE 4 <u>MA. 1 Microfacies Association.</u>

A Calcisphere wackestone / packstone. MA.1.1. s/m 139, Tynant Formation, World's End, G. R. SJ 23304788

B Unlaminated calcisphere peloid grainstone. MA.1.2a s/m 1003, Tynant Formation, G.R. SJ 22064588.

C Vague peloids visible in very thin section (feather edge). Note diffuse margins and micrite/cement matrix. s/m 1590, Froncysyllte, Tynant Formation, G. R. SJ 26984199

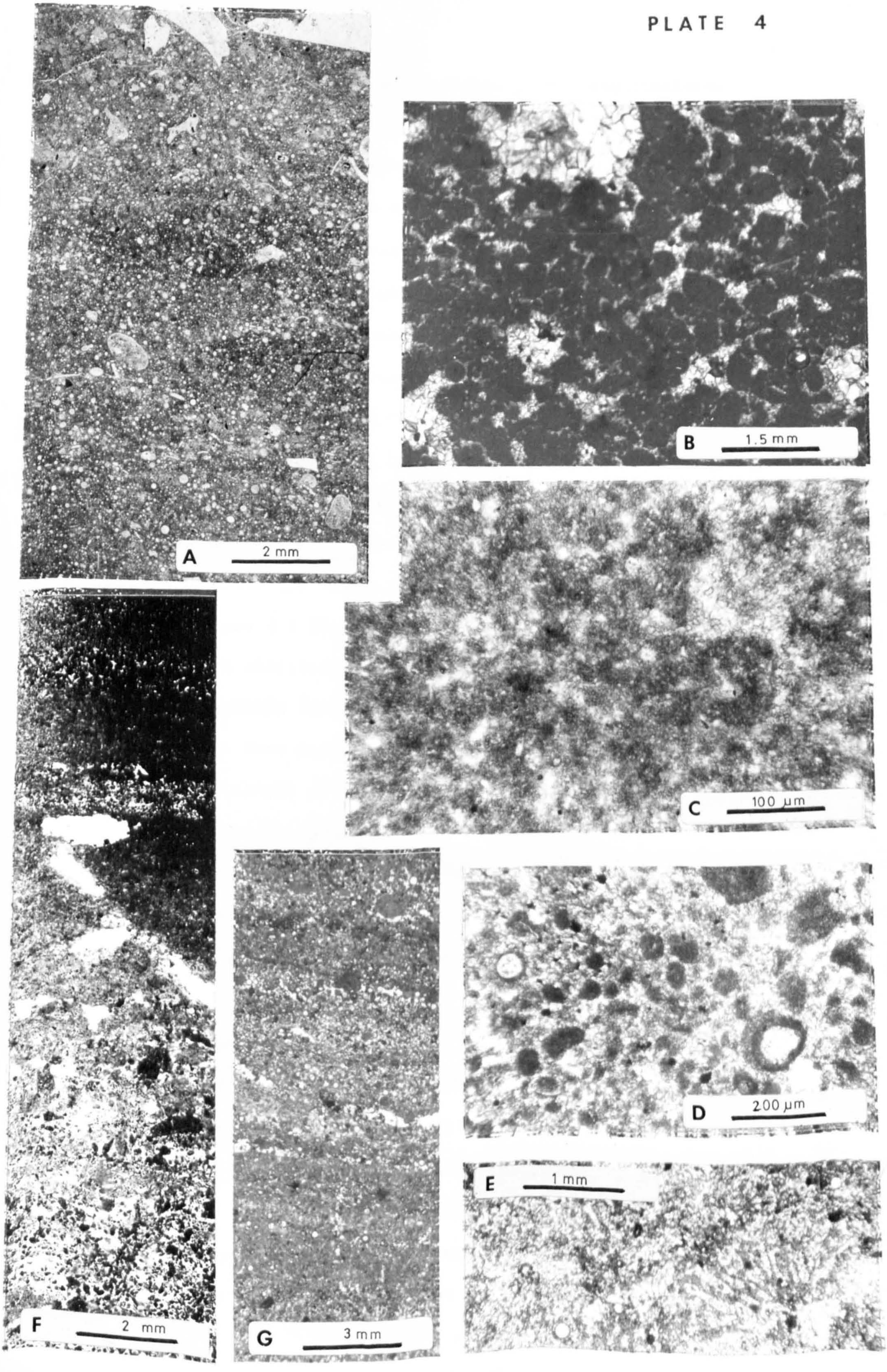
D Well defined peloids in MA.1, s/m 246ii, Tynant Formation, Tynant, G. R. SJ 21984572

E <u>Garwoodia</u> sp 'colony' within MA.1.1., s/m 158, Tynant Formation, World's End, G. R. SJ. 23284819

F. Laminated calcisphere peloid grainstone /packstone MA.1.2b, Tynant Formation, G. R. SJ. 21984572.

G Vague grading in millimetre lamination of MA.1.2b, s/m 246, Tynant Formation.

PLATE 4



(1977, p55) described similar recent peloid associations.

The calcispheres are of many genera and species, although in the Sychtyn Member and Tynant Formations the very thick-walled highly spinose parathuramminids are especially common (figs 11 & 12). Of the 'smooth walled' calcispheres, Pachysphaera spp.are very common. Gastropods are the dominant macrofaunal element,both high and low spired forms. Their shells are always replaced by coarse calcite druse and occasionally have included internal sediments and void-collapse features. The proportion of mollusc clasts within the sediment (both whole and fragmentary) is variable, but not greater than other microfacies, although the exclusion of other macrofaunal elements increase their significance within MA.1. Recognisable bivalve clasts are rare. Brachiopod, echinoderm and coral bioclasts are also rare (< 1% of volume), but when present they are invariably abraded and micritised.

Beresellids (mainly Kamaenella sp .) are volumetrically less significant than calcispheres, but with their proportional increase, along with clasts of 'stenohaline' elements, MA.1.1 grades into MA.2.1. Ostracods are ubiquitous and may occasionally be the most important sediment contributor. Foraminifera are minor sediment contributors. Dasycladacean and filamentous algae are rare, as are ungdarellids (e.g. s/m 197). Occasional lamina-dependant Garwoodia and Ortonella (e.g. s/m 158 and 188) masses and cryptalgal laminations occur (Plate 4, fig. E).

Bioturbation other than tubular fenestrae and sutured discontinuity surface (Chapter 11) features is rare, usually evident as horizontal and vertical pipes infilled with more mud-deficient packstone-grainstone sediment with variable quantities of geopetal silt. It is these burrow-fills that suggest the peloid nature of the surrounding 'lime mud'. At two horizons in the Sychtyn Member (G.R. S.J. 25102505) Chondrites sp.† occurs towards the top of a

† Plate.1, fig F

wackestone unit, but with burrows infilled by overlying sediment. Appendix II lists the compositional range of MA.1.1.

6.2.2. Petrography of MA.1.2a. Unlaminated calcisphere-peloid grainstone/packstone (Plate 4, fig B)

MA.1.2a is commonly interbedded with MA.1.1. It normally has fenestral fabrics associated, usually a combination of small and medium irregular, laminoid, tubular and 'packing' fenestrae. Large irregular fenestrae are rare. The laminoid and 'packing' fenestrae impart or enhance a grainstone texture, even in sediments that are apparently packstone. Clast size is variable between 50μm and 1000μm.

Most peloids are subrounded to irregular with a micritic internal structure, but often have a well developed coating of dense micrite visible as a pale rind on polished surfaces.† Apart from calcispheres, recognisable clasts are few. Intraclasts of calcisphere wackestone are present, along with minor percentages of foraminifera, ostracods, molluscs and echinoderm fragments.

6.2.3. Petrography of MA.1.2b. Laminated calcisphere-peloid grainstone-packstone (Plate 4, fig. F).

MA.1.2b is a rare microfacies. It is best exposed in the Tynant Limestone (G.R. SJ 21954570, Chart A, s/m 246) as a 20cm bed, laterally traceable over c. 200m. It comprises 500μm to 2000μm laminae of calcisphere peloid grain-supported sediments with rounded elliptical intraclasts ('flakes' in '3D') up to 1cm long. Detrital quartz is abundant along some laminae. The thicker (2mm) laminae show vague fining upwards grading (Plate 4, fig. G) with lime-mud concentrated towards lamina tops. Small laminoid fenestrae are ubiquitous with small irregular and packing fenestrae subordinate. Cryptalgal laminites (MA.1.3.) are intimately associated with this microfacies.

At the Tynant locality the microfacies is layered on a milli-

† pisolites - see section 6.4.1.

metre to centimetre scale, subparallel with but not totally dependent on the sediment lamination, between partly dolomitised and non-dolomitic layers. Its distribution and the presence of dolomitic intraclasts within non-dolomitised laminae show that dolomitisation was early. This is described in section 6.5.8.

6.2.4. Petrography of MA.1.3. Cryptalgal laminite (Plate 5, fig. A).

Cryptalgal laminites occur within MA.1 associations in minor quantities as scattered millimetre to 10cm thick layers. They have pale dense micritic laminae 50 to 300µm thick. Sometimes the laminate character is very vague, defined by relative proportions of included detrital quartz (~ 20µm) or by small laminoid fenestrae, or by peloid-calcisphere laminae. The laminae may be partly disrupted (e.g. s/m 2109) and fragmented (Fig. 38) into elongate flat clasts. This disruption indicates their early lithification (see section 6.5.7.). Specimen 2155 from Treflach Quarry in the Sychtyn Member shows ?plastic deformation by vertical zones of intense tubular fenestration upturning but not fracturing the laminae (Plate 5, fig. D).

Microrelief is visible on some cryptalgal laminites, although no large-scale stromatolitic structures have been observed. Low, (2cm relief) 8cm to 12cm base, smooth domal forms ("LLH" of Logan et al 1964) occur along one horizon within MA.1.1 in the Sychtyn Member (fig. 39). The Tynant Limestone (G.R.SJ 24685113, fig. 40) shows laterally impersistent cryptalgal laminites with 20cm primary relief, thin laterally extended sheets, and low domal LLH forms.

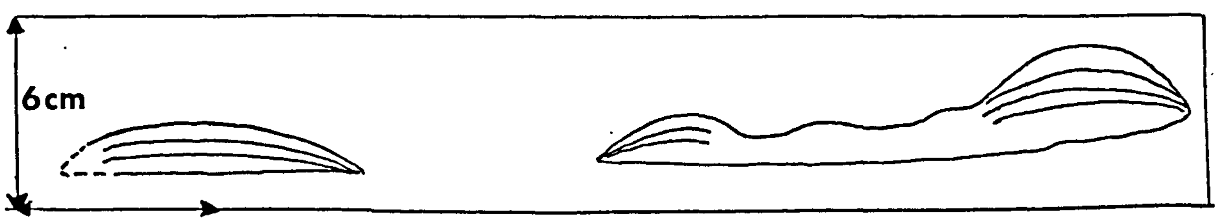
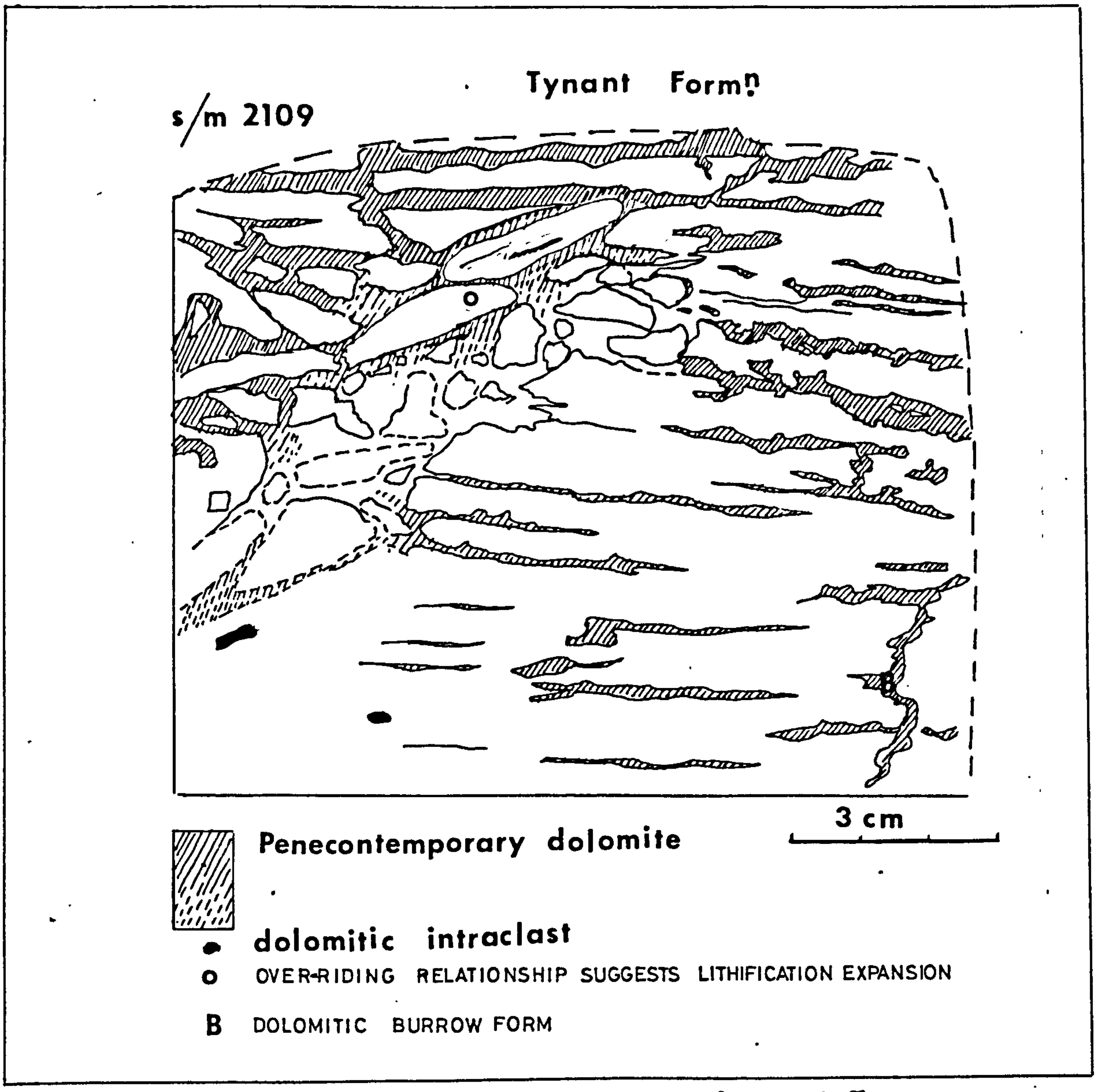


Fig. 39 Stromatolitic (LLH) structures within MA.1 of the Sychtyn Member, Nant Mawr Quarry, G.R. SJ 25232500.



Early lithification and <u>in situ</u> fracture of MA I. 3

Diagram of lateral variation in regressive phase microfacies of a Tynant Formation minor cycle. g.r. SJ 24685113

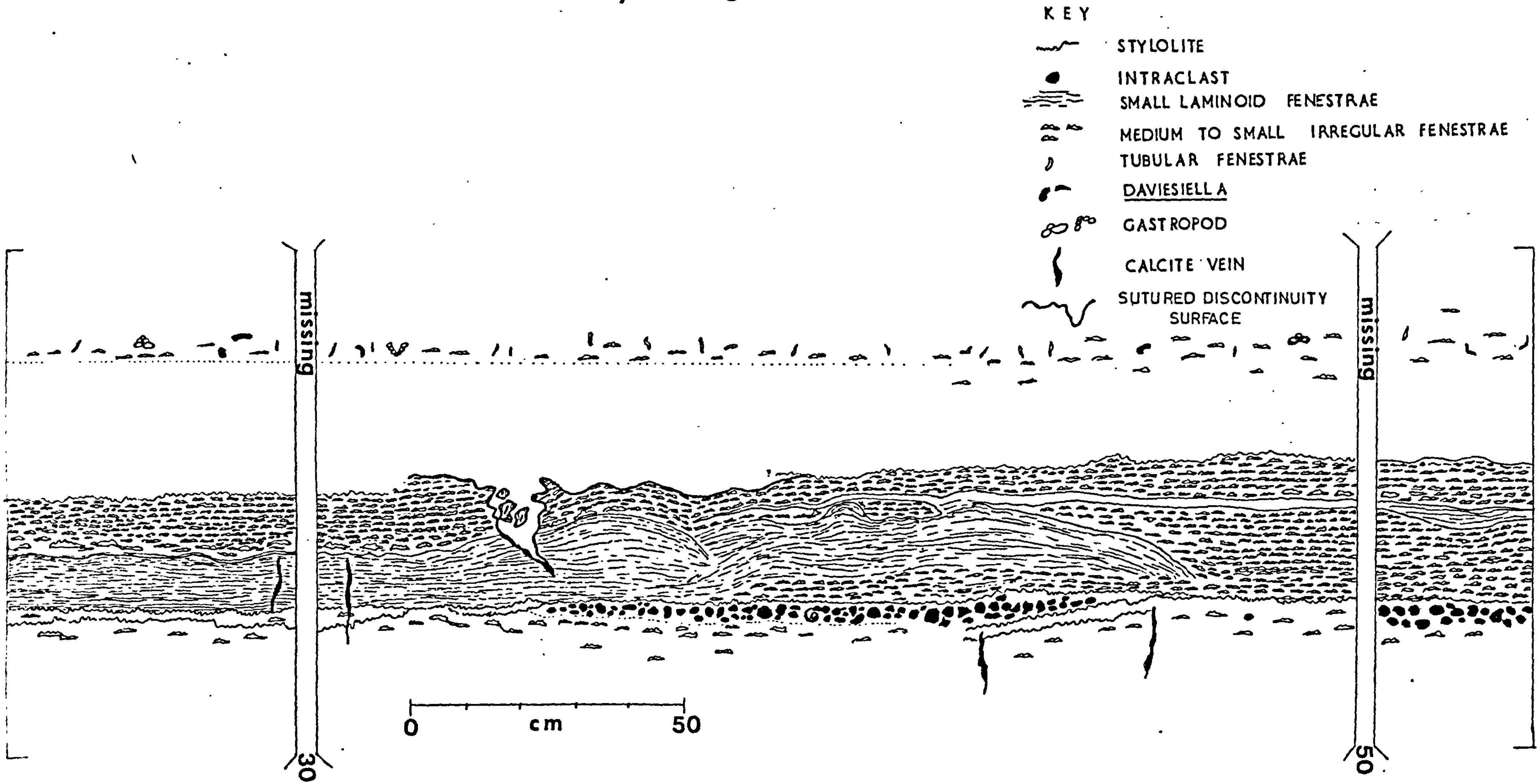


Fig 40

PLATE 5

A MA.1.3 micrite-detrital cryptalgal laminite ?quartz rich laminae with fine laminoid fenestrae s/m , Tynant Formation, G. R. SJ

B Laminated beresellid calcisphere packstone-mudstones MA.1.4 s/m 1101, Sychtyn Member, G. R. SJ 25252500

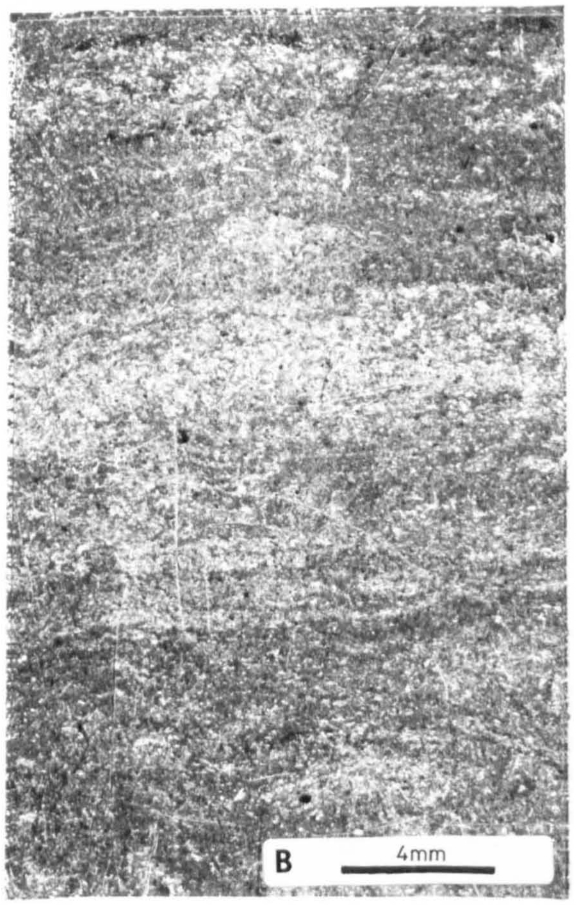
C Fenestra with internal sediment of pellet grainstone. Note tubular fenestra (R) within the internal sediment S/m 1498, Tynant Formation, G. R. SJ 22014552.

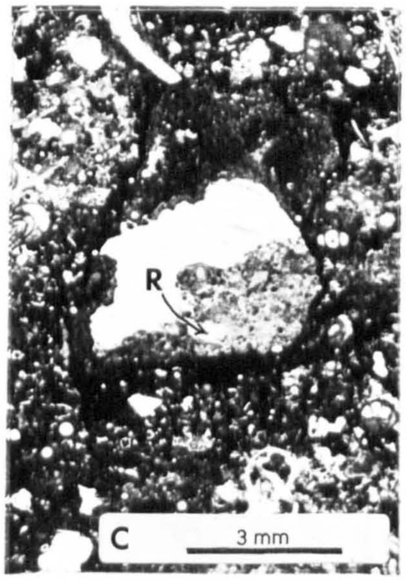
D Intense tubular fenestration within MA.1.3, s/m 2155, Sychtyn Member, G.R. SJ 25252499.

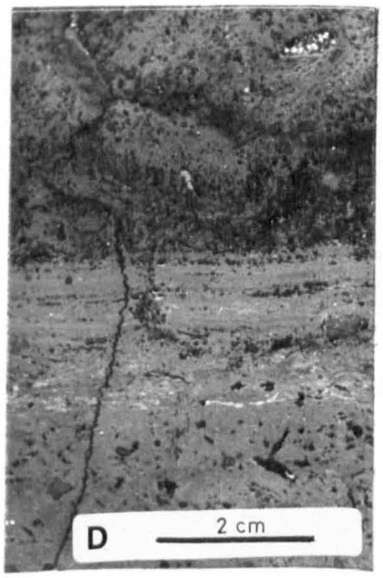
E Rhythms of intense tubular fenestration as in (D), Sychtyn Member.

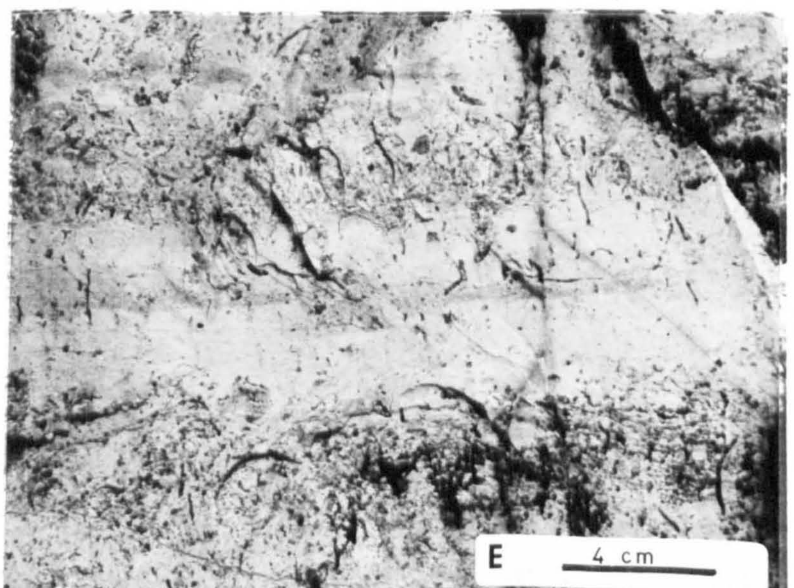
F Refenestration of internal sediments within an earlier fenestral feature., S m 289. Tynant Formation, G.R.SJ21994551. (original fenestral fabric outlined).

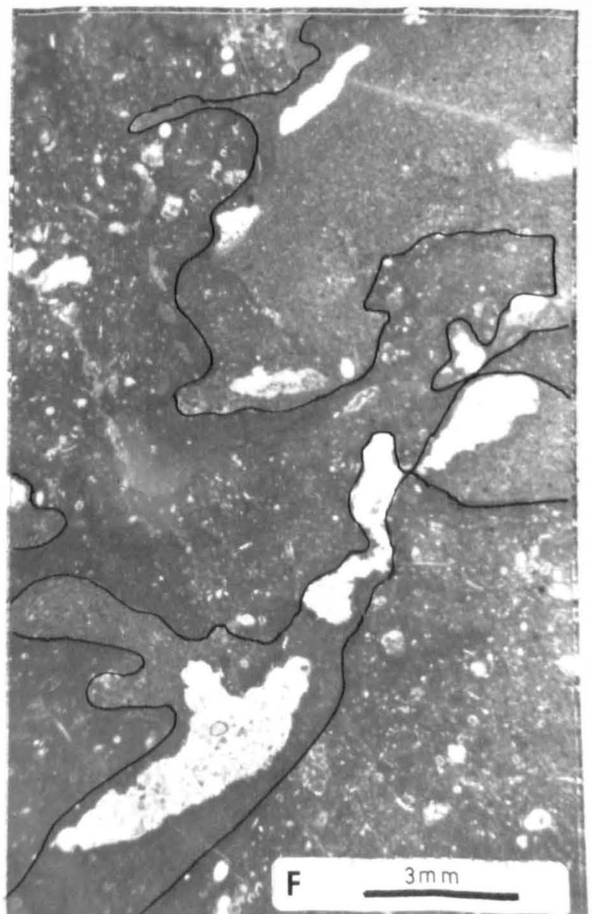
PLATE 5











6.2.5. Petrography of MA.1.4. Laminated beresellid-calcisphere mudstone-packstone. (Plate 5, fig. B).

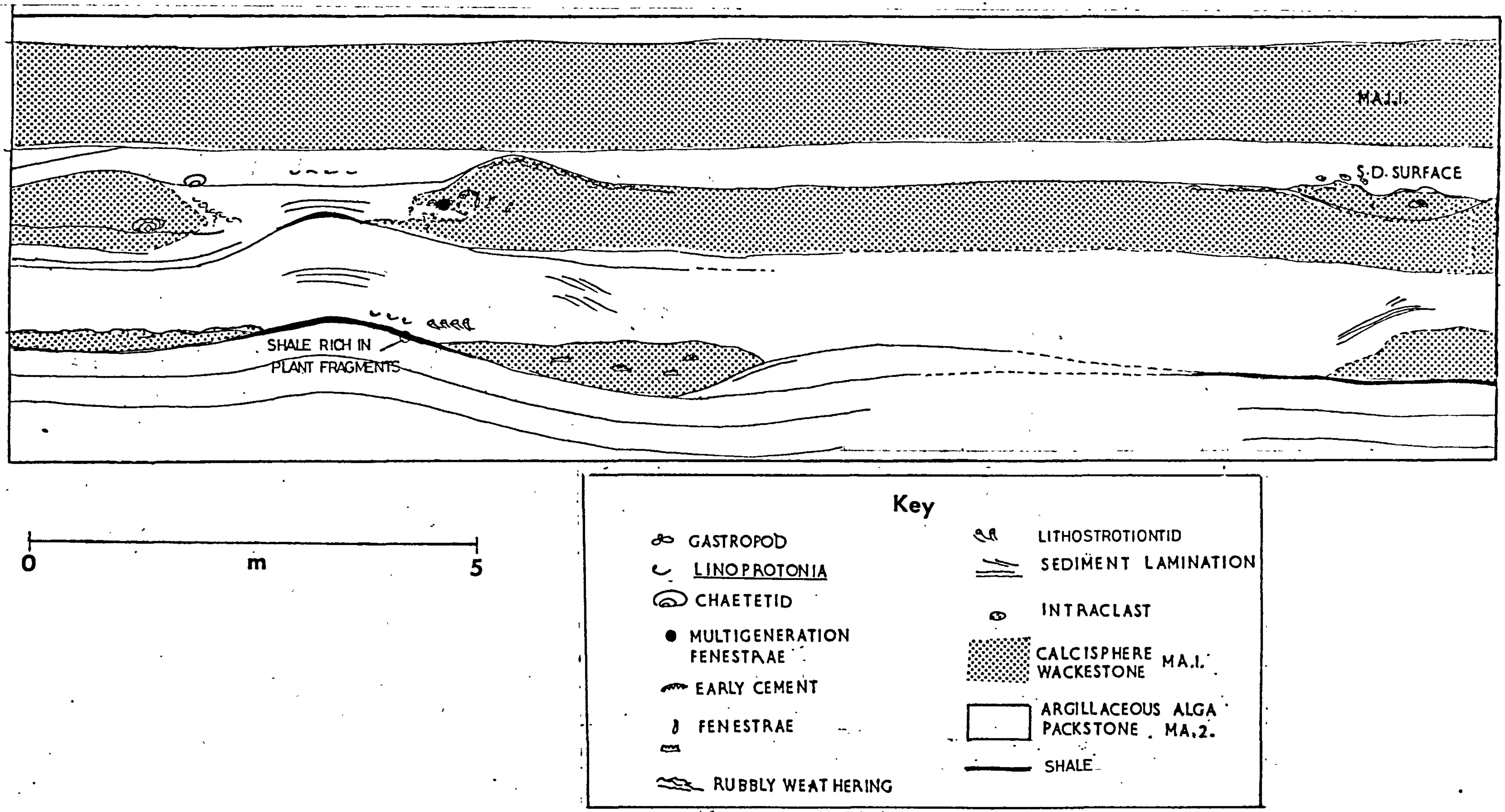
Present in the regressive phases of Sychtyn Member cycles and rarely in the Tynant Formation (e.g. s/m 239.) this microfacies is compositionally a transition lithology between MA.1 and MA.2 with proportions of beresellids ≃ calcispheres. It occurs as units ≤30cm .S/m 1101(Sychtyn Member) has vague to prominent, slightly crenulate millimetre lamination, sometimes with a superimposed darker-lighter brown colour layering on a millimetre to centimetre scale. This colour banding is laterally variable and gives rise to 'stranded' dark colour mottles. (e.g. s/m. 1101). The lamination is between beresellid (Kamaenella sp.) mudstones to packstones, and more calcisphere-peloidal grainstones similar to MA.1.2b. Unlaminated MA.1.4. also occurs rarely.

Stenohaline faunal elements are absent. Rare bioclasts of echinoid spines, Koninckopora sp. and endothyrid foraminifera occur.

6.2.6. Character of Clast Supported Conglomerates.(L.7.1)

This Lithofacies L. 7 is characteristically associated with MA.1 sediments. It occurs above irregular erosion surfaces within MA.1 units. Clasts are almost totally MA.1 sediments, and range up to 8cm diameter (Fig. 41). They are normally irregular to subspherical, and little rounded. Matrix is characteristically lime-mud and intraclast and none have been observed with imbricate structure . The dominant texture of the larger intraclasts is clast-support, although this does grade to matrix supported, especially towards upper zones of the conglomerate 'bed', which rarely exceeds 15cm thick. No 'flake breccias' have been observed.

Fig 41 Bed geometry of Tynant Formation at Tynant Quarry SJ 21964573



6.3. BED MORPHOLOGIES OF MA.1.

Small-scale laterally discontinuous units of MA.1 sediments are visible in the Tynant Limestone at outcrop whilst more regional variation is discovered from correlative comparison of logs (refer to charts A and C).

At Tynant Quarry (G.R.SJ21944560) and Trefor Rocks (G.R.SJ22664348) the Tynant Formation shows a number of pinch and swell features (Fig. 41) within MA.1. These are: 1) The development of a channel-like body 3m wide and 80cm deep infilled with MA.1.1 with a coarse irregular fenestral fabric, that apparently extends ENE into the quarry face (Plate 1, fig. A): 2) relatively abrupt lateral termination of a bed: 3) possible channel form within MA.1 infilled by MA.2. Somerville (1979a,p.401) has interpreted the former two features as palaeokarst on top of his 'cycle 10'. The lack of karstic potting and the preservation of S.D. surface features suggest karstification was a minor process. The presence of multigeneration fenestrae (fig.41 ;Plate 5, fig.C&F) imply a later reactivation of the sediments after early cementation, but whilst being only poorly lithified. The bed irregularities are related to undulations in the underlying MA.2 sediments and may be in-part an accretionary phenomenon on a primary relief. Whether this relief is of tectonic or depositional origin is unclear. There is no evidence of fabric termination, nor primary accretion surfaces on the 'erosive' margins of these bodies. The presence of immediately overlying intraclastic layers implies that these bed morphologies were in part modified during the succeeding transgressive event.

The top Tynant Formation cycle also displays a laterally discontinuous bed morphology that shows internal fabric truncation† (fig. 9 ,G.R.SJ22204668) indicating that erosion plays an important

† see section 11.1

role in their formation (see section 6.4.2.5 & 11.1.2).

In microtidal (Davies 1964) shorelines (see section 6. 7.) reduced tidal currents inhibit the formation of tidal channels and creeks, although the extent of tidal flat development (also controlled by seaward gradient) is less for lower tidal ranges. Read (1974) described small-scale intertidal creeks in Shark Bay, Western Australia. Intertidal channels migrate laterally (Shinn et al., 1969, p.1221), building levees and fining upward units. There is no evidence of these features in the Tynant Formation. This does not preclude their interpretation as minor low-energy and short lived channel-forms, although karstic removal† and erosional scour appears a more acceptable model. † (including intertidal dissolution - see section 11. 1)

Topographic relief formed by karstification is described in chapter 11 , and the larger scale embayment-wide variations in MA.1 geometry in chapter 2 .

6.4. EARLY DIAGENETIC AND SYNSEDIMENTARY FEATURES ASSOCIATED WITH THE PERITIDAL SEDIMENT SUITE

6.4.1. Pisolitic, pseudooolitic & oncolitic clasts

Subspherical grains with a dense pale micrite coat $\leqslant 4$mm thick to clasts, up to 4cm diameter, are common within MA.1.2 sediments, intraclast rudstones and intraformational conglomerates associated with MA.1 microfacies.

These structures lack calcareous algal filaments and concentric lamination. The coats are irregular in thickness on individual grains but neither'vadose dripstone' and deformation features (cf. Dunham, 1969b) nor thickening on clast tops (Hay and Reeder, 1978) are apparent. Coats thicken, apparently at random points. These coats resemble the pale dense early micrite cements of fenestral cavities (see on) and lack the dark tan pigment colouration of calcrete micrite coated clasts. No'coarsening upward' fabrics have been recognised (Dunham 1969b).

The coats differ from MA.3 oncolitic clasts in lacking porous (fenestral) fabrics and calcareous algal filaments (see section 7. 3. 2).

An in part accretionary origin for this micrite coating is suggested, both by the sharp boundary on many (but not all) grains between 'coat' and host 'clast(' and by the presence of ?included silt grade material within the coats. (e.g. s/m 1494, Tynant Formation, fig. 8, Plate 9, fig, D.) .

Although Read (1974b) described "calcrete ooids" and pisolites from Shark Bay soil zones lacking the meniscus and packing effects of Dunham (1969b), the lack of pigmented calcrete micrite and other calcrete structures associated precludes a soil zone calcrete interpretation. (Esteban, 1976, refuted the soil process formation of Dunham for pisolites of the Capitan reef complex, suggesting instead a syndepositional algal origin.)

The accretionary origin of these coated clasts distinguishes them from the endolithic algal-micritisation process of Bathurst (1966), although it retains a degree of similarity with the exolithic algal micritisation described by Kobluck and Risk (1977). Non-skeletal oncolites are common tidal-flat associated cryptalgal structures in both the Recent and Ancient. Comparable 'oncolites' to these Lower Carboniferous forms have been recorded from peritidal algal mat zones of Shark Bay (Logan et al, 1974a,p. 152 as "pelletoid rinds") and the Jurassic (Purser, 1975). Logan et al (1974a) interpret their aragonitic pelletoid rinds as products of algal trapping processes, a model favoured here for these Lower Carboniferous analogues.

The presence of cryptalgal pisolites suggests formation in an environment of active or ephemeral algal growth on inter-tidal flats.

6.4.2. Fenestral Fabrics.

The classification followed here is a modified form from Logan (1974, p.214).

6.4.2.1. LAMINOID FENESTRAE Small laminoid fenestrae are associated with algal, cryptalgal and peloidal laminites (MA.1.2a and b ; MA.1.3). They enhance any grainstone character of the sediment, and are themselves modified by the grain shape and packing. They sub-parallel internal sedimentary lamination, are between 100μm and 1000μm tall and can often be traced for up to 4cm laterally (Plate 6, fig. C). Tubular fenestrae often interconnect with them. Lateral to the fenestrae evidence of disruption is shown by internal fracturing and roof spall floating in geopetal silts.

Floors are generally smoother than ceilings: in lime-mud rich textures, both floor and ceiling may be smooth and parallel, whereas grain-support textures with internal geopetal sediments have fenestrae with more irregular ceilings and smooth floors. With decreasing size and more irregular form they grade into 'packing fenestrae' type.

Fenestra density is variable, but normally depends on the scale of sediment lamination. More finely laminated fabrics have a denser fenestral distribution.

Medium to coarse laminoid fenestrae are rarely developed (e.g. Tynant Formation at, G.R.SJ21944560). They are greater than 1mm tall and often extend laterally more than 4cm. Coarse sub-laminoid forms up to 3cm long are here classified as irregular fenestrae

These fenestrae subparallel sediment lamination, and are rarely inclined at a shallow angle (Plate 6, fig. B). They merge with intergranular and umbrella voids. Internal sediments are of variable quantity, but as with smaller laminoid fenestrae, they are generally reduced compared with other fenestral fabrics that have a more dominant vertical component.

6.4.2.2. IRREGULAR FENESTRAE. With increased vertical dimension relative to a reduced lateral extent, laminoid fenestrae grade into 'irregular' forms.

Small irregular fenestrae are subcircular to vertically compressed ellipses, with irregular margins, ca. 0·1–4mm tall. This is the most common fenestral fabric, occurring in units up to 150cm thick, mostly in MA.1.1 and MA.1.2a. These vary from dense ($\gtrsim$ 30% by volume) to ($\lesssim$ 5% by volume) scattered. They grade into medium irregular fenestrae, and are normally associated with these, tubular

PLATE 6 Fenestral fabrics.

A Medium to small irregular fenestrae. Note vague concave-up fenestral pattern outlined. S/m 1472, Sychtyn Member, Craig y Rhiw G. R. SJ 236295.

B Medium to coarse laminoid fenestrae, in MA.1.2 S/m 1003, Tynant Formation, G. R. SJ 22064588

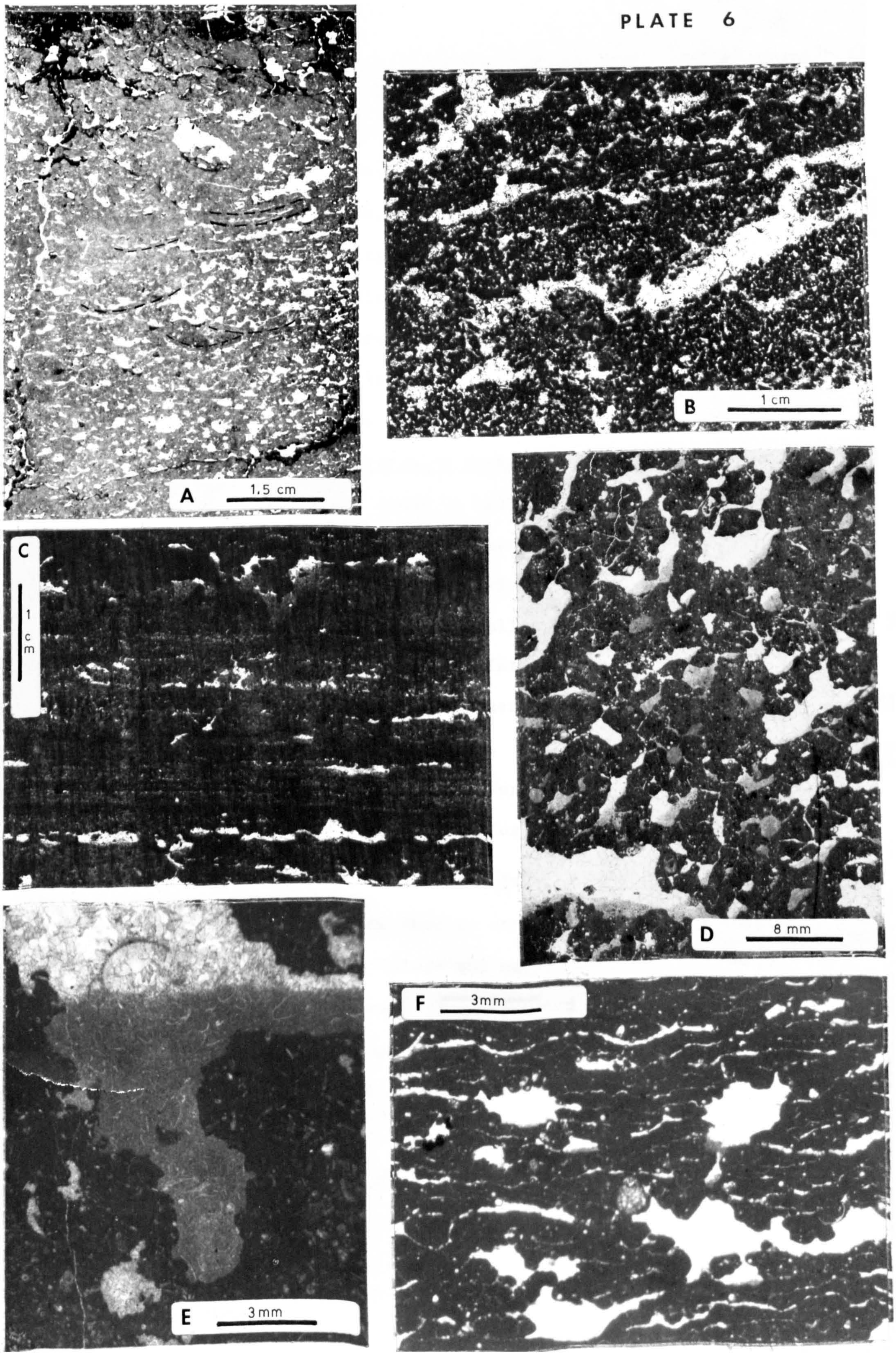
C Small laminoid fenestrae in MA.1 2b s/m 246. Tynant Formation, Tynant.

D Irregular 'packing fenestrae' extending between medium to small irregular fenestrae defining ped-like pseudointraclasts. S/m 130, Sychtyn Member, Nant Mawr Quarry.

E Irregular fenestral cavity within MA.1.1 infilled by muds and silts rich in ostracod carapaces. S/m 1456, Leete Formation (Mesothem D.5a). Llanelidan, Pen y Graig, G. R. SJ 11655190.

F Small irregular fenestrae with sub-laminoid packing fenestrae. S/m 1098 Sychtyn Member, Nant Mawr Quarry.

PLATE 6



and packing fenestrae. Grainstone textures produce more irregularly shaped fenestrae. Internal sediments are variable but only rarely completely fill fenestrae. Tubular fenestrae often interconnect in all directions, and packing fenestrae merge with the irregular fenestrae giving their edges a fine 'root-like' structure penetrating the surrounding sediment.† Indistinct stacked concave-up subhorizontal (2cm) trains of fenestrae commonly occur, producing a scalloped laminar texture, probably defining original sediment microrelief (see section 6.4.5.3 , Plate 6, fig. A).

Medium and coarse irregular fenestrae are $\geqslant$ 4mm, and occasionally > 10mm tall. They occur in both MA.1.1 and in more grain-supported textures, sometimes as a unit of limited lateral extent. They are usually scattered and comprise $\leqslant$ 10% by volume of rock. Medium irregular fenestrae are normally associated with small irregular and packing fenestrae. Internal sediments are common and variable, sometimes showing sequential infill laminae with interlaminated early cement fringes. Vadose cements occur in these fenestrae (see p.104). Morphologically, these fenestral fabrics grade into coarse laminoid forms.

6.4.2.3. TUBULAR FENESTRAE. Tubular fenestrae are ubiquitous to all the MA.1 suite. They are tubular structures 100 to 2000µm diameter, with smooth, undulose and subparallel to highly irregular outer surfaces. They have no wall structure (cf tubular palaeosol structures, p.231). They disperse in both vertical and lateral components through the sediment, coalescing with other fenestral fabrics when present, and occur in variable density, from scattered to being often the only or dominant fenestra type. One regressive phase in the Sychtyn Member (s/m 2155) has three 3cm layers densely packed with these fenestrae ($\geqslant$ 30% by volume) separated by 4cm to

† Plate 6, fig. A & D

8cm sparsely tubular fenestrated MA.1.1, and cryptalgal fabrics (Plate 5 fig. D&E). They branch and coalesce in all directions, following existing fenestral fabrics where present by offshoot. In laminated sediments horizontal components may dominate, giving these fenestrae a laminoid appearance.

Internal sediments are common, often occurring as pyramidal heaps on horizontal floors beneath vertical components of the fenestrae.

6.4.2.4. PACKING FENESTRAE. This is a new term, defined here as a minor (by volume) but common fenestral type, enhancing intergranular voids, floating clasts, and in matrix-rich sediments forming a reticulate network that imparts a pseudograinstone texture to the sediment (Plate 6, fig. D). These fenestrae 'fractures' follow margins of included clasts. With increased horizontal component they grade into small laminoid fenestrae (Plate 6, fig. F).

Their distribution is widespread, occurring in association with all other fenestral fabrics (excepting large irregular fenestrae), but are especially important in association with small laminoid and small irregular fenestrae.

The fenestrae are from 10μm to 100μm wide, extending in all directions in a reticulate manner (in cross-section). They have irregular margins, often widen towards other fenestral features and at confluences but also can fade into the surrounding sediment. As with all other fenestral fabrics they have variable quantities of internal sediments, concentrated on horizontal bottom surfaces that underlie vertical elements of fenestrae.

Grover and Read (1978) included these packing fenestrae within their laminoid fenestral group (op. cit. p.460, fig.D). Deelman (1972, p.590) illustrated the formation of enlarged voids by upward migration of air bubbles displacing sediment particles thereby enhancing the grain-support texture to grain- float. This, and desiccation shrinkage, may be an important mechanism for the formation of packing fenestrae.

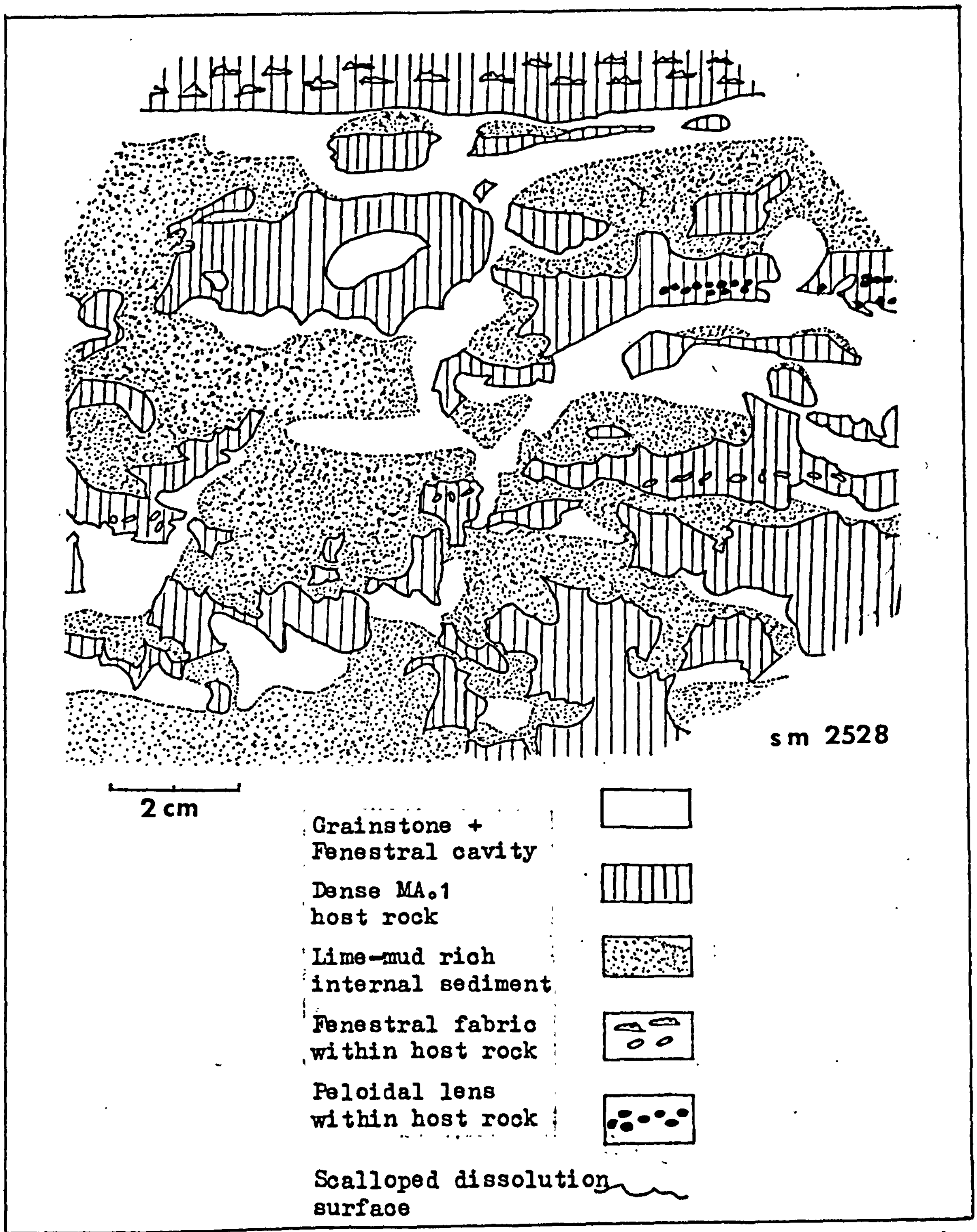
6.4.2.5. MULTIGENERATION FENESTRAE. At a few localities in the Tynant Formation two successive generations of fenestrae occur within the same unit, superimposed upon each other. In s/m: 027; 289 and 1498 small and medium irregular fenestrae are infilled with peloidal-bioclastic silts and mud, that has been penetrated subsequently by tubular fenestrae (Plate 5, fig. F). Specimen 2528 has small irregular and tubular fenestrae, partly infilled with bioclastic-silt, mud and micrite cement. Subsequently lithified, this sediment has been partly dissolved (section 11.1.2) producing a distinctive, large scale fenestral fabric with mm scalloped margins. (fig. 42), themselves infilled by sediment subsequently ?burrow-fenestrated.

The former examples indicate sequential sediment modification, and show that tubular fenestration may be a later event to irregular fenestral fabrics. However, these specimens immediately underlie cycle boundaries and refenestration most probably occurred during the succeeding transgressive phase. The succeeding section continues this argument from a superposition viewpoint.

6.4.3. <u>Discussion on Fenestrae.</u>

Ham (1952) introduced the term 'birdseye' to describe calcite-spar druse-infilled irregular vughs. Tebbut <u>et.al</u>. (1965) used the more embracing term 'fenestra', which has been followed more rigorously by subsequent workers. Fenestral features are now powerful peritidal palaeoenvironment indicators, but must be used cautiously within an assemblage of criteria to define an emergent environment (Ginsburg 1975).

Ham (1952) proposed their formation from decay of algal remains, whilst Illing (1959) suggested that they have many origins



Dissolution fenestrae associated with minor cycle boundary in Tynant Formation (fig 9b) including organic borings, syneresis and desiccation shrinkage, and gas evolution from organic decay. Cloud (1960) used the presence of pyrite within fenestrae to argue their gaseous formation, whereas Deelman (1972) used a capillary water transport and gas-bubble (entrapped air) escape mechanism to account for their origin.

From his study of alpine Triassic 'Loferites' Fischer (1964) proposed their intertidal origin. Shinn (1968) confirmed this from investigation of Recent supratidal sediments and formed 'birdseyes' in the laboratory by simulating alternate wetting and drying, but in alga-free lime mud. Logan (1974) and Logan, Hoffman and Gebelein (1974) studied the variation and association of fenestra types in Shark Bay peritidal sediments, and recognised their formation due mainly to interaction between algal-mat and algal-bound sediment, but desiccation, (watertable fluctuations) oxidation (Eh/pH fluctuations) and lithification also are important (Logan, 1974, p.214). Fenestrae can form up to 30cm below the sediment surface. Three inter-gradational categories of fenestra were erected by Logan (*op. cit.*) that are followed here, *viz* laminoid, irregular and tubular morphological forms.

Grover and Read (1978) presented a comprehensive study of fenestral types, densities and distributions in Ordovician carbonates from Virginia. Read (1975) described Devonian tidal-flat sediments from Western Australia, and noted a sequential arrangement of fenestral fabrics within cycles that he compared to the sub-environmental distribution of Shark Bay fenestral fabrics.

6.4.4. Sequential Distribution of Fenestral fabrics in the Tynant Formation and Sychtyn Member

Fenestral fabrics occur in many but not all MA.1. units. At times they are highly variable in vertical association and grade laterally over metres of outcrop into other or non-fenestral fabrics.

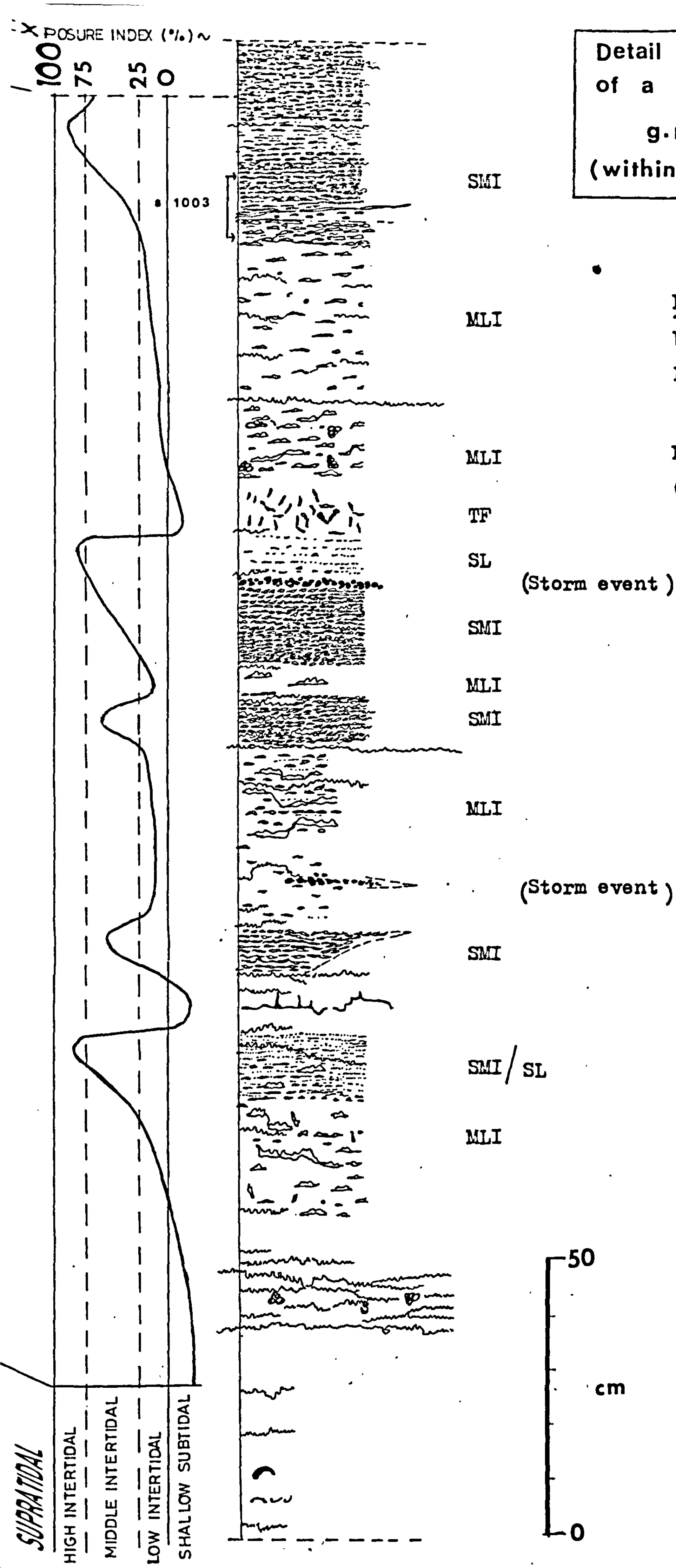
Fig.43 indicates the degree of vertical variation encountered.

To evaluate whether these lower Asbian fenestral fabrics may be of use as palaeoenvironment indicators, and whether they have a prefered vertical (cyclic) distribution, a Markov analysis was undertaken:

A Markovian process is one in which the likelihood of an individual event is partly determined by previous event(s). Many analytical tests for Markovian processes are available and usage depends upon measured characters, data-set size, and aims of study.

Neither individual section thickness nor exposure quality were sufficient to obtain statistically significant data sets. Combination of data from a number of Asbian sections, mostly within the upper series of regressive-phase-dominated minor cycles of Mesothem 'D5a' produced a single data set of 124 dominant fenestral fabric transitions. This set includes 'multistorey' transitions (<u>sensu</u> Selley, 1970) recognised by changes in fenestra density, bedding planes and intraformational conglomerates. As the population is limited a simple statistic was chosen (i.e. the method of Selley, 1970), the data did not permit tests of 'stationarity' (e.g. Casshyap (1975)). Only tests for 'first order ' Markovian processes (Schwarzacher, 1975) were attempted.

A 'tally' matrix was produced to indicate the frequency of fabric transitions within the data set. Data from different



Detail of the 'regressive' phase of a Tynant Formation cycle.

g.r. SJ 22064588

(within MA 1 microfacies)

Fig 43

Estimate of 'exposure' based on analogies with Recent fenestral fabrics (section 6.4.5.).

Exposure Index concept after Ginsburg and Hardie (1975)

Key

stylolite

small, large } fenestrae

'Sheet cracks'

gastropods

Daviesiella

large intraclasts (>5mm)

tubular burrow fenestrae

SMI small - medium irregular fenestrae

M LI large - medium irregular fenestrae

TF tubular fenestrae

SL small laminoid fenestrae

stratigraphic sections were 'run-together' by including 'non-fenestrated' sediments as shared events between stratigraphic sections ending and commencing (9 sections in toto). Using Selley's Method (see Schwarzacher, 1975, for statistical theory) a data matrix was constructed for predicted random transitions if no Markovian process was acting on the fenestral fabrics. Gingerich's (1969) method was also used as a comparative.

The resultant 'non random' transitions found by subtraction of the random matrix from the observed tally of fenestral fabric transitions are plotted in figs. 44 & 45.

Application of a simple 'X^2' statistic (Schwarzacher, 1975) to this data set indicates that at 'X^2' = 34.4 and 25 degrees of freedom, the 'null hypothesis' may be rejected with 90 - 95% confidence. i.e. There is less than 10% chance that such a distribution of probabilities could arrise from successive independent events. The paucity of the data does, however, suggest caution must be employed in drawing conclusions.

As fenestral fabrics are in part controlled by substrate type, and morphology of algal mats (Logan, 1974) if present, they may be reasonably expected to reflect tidal flat subenvironments.

The fabric relationship diagrams promote discussion: A general cyclic pattern occurs from tubular or large irregular fenestrae through smaller irregular to laminoid fenestrae.

In comparison to the apparent distribution of these Lower Carboniferous fenestral fabrics, Read (1973, 1975) described vertical successions of fenestral fabrics from the Devonian of Western Australia. He recognised a general upward sequence from non-fenestrated pellet limestones through laminoid and irregular fenestral fabrics into pellet limestone

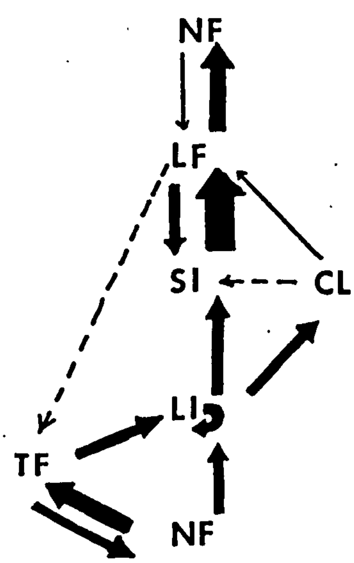
Fig. 44 Matrix showing the difference between the observed number of transitions from one fenestral fabric to another in an upward succession, and those predicted assuming a random model for their successive formation. (Selley's method)

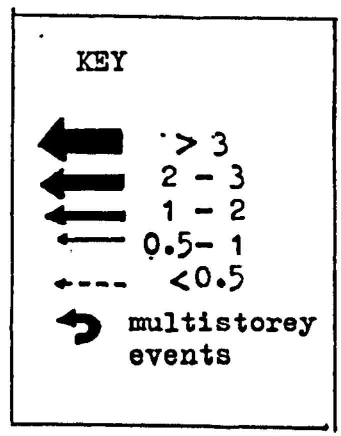
(see fig.44a for tally matrix and random matrix)

		LF	SI	LI	TF	CL	NF
Laminoid fenestrae	(LF)	-1.87	+1.82	-2.87	+0.49	-0.30	+2.70
Small to med. irreg. fen.	(SI)	+3.82	-0.34	-1.18	-1.25	-0.65	-0.39
Medium to lg. irreg. fen.	(LI)	-2.87	+1.82	+1.13	-0.51	+1.70	-1.28
Tubular fen	(TF)	-1.51	-1.25	+1.49	-0.79	-0.16	+1.22
Non-fenestral cryptalgalaminite	(CL)	+0.70	+0.35	-0.30	-0.16	-0.03	-0.56
Non-fenestrated sediment.	(NF)	+0.70	-2.39	+1.72	+2.22	-0.56	-1.72

'χ^2' =34.4. ; $(6-1)^2$ degrees of freedom : i.e. The null hypothesis can be rejected with 90 - 95% confidence.

Fenestral fabric relationship diagram from the above matrix showing the upward transitions which occur most frequently for each fenestral fabric, after allowing for the number expected, had their formation been random.





From a population of 126 fenestral fabric association transitions.

MATRICES FOR MARKOVIAN ANALYSIS

1) SELLEY'S METHOD.

A) TALLY MATRIX

	LF	SI	LI	TF	CL	NF
Laminoid fenestrae (LF)	1	8	0	2	0	8
Small irregular fenestrae (SI)	10	13	5	2	0	11
Medium -large irregular fenestrae (LI)	0	8	4	1	2	4
Tubular fenestrae (TF)	1	2	3	0	0	4
Non-fenestrated cryptalgal laminite (CL)	1	1	0	0	0	0
Non fenestrated, non-laminated (NF)	6	9	7	5	0	8

Total of 126 transitions

B) RANDOM MATRIX

	LF	SI	LI	TF	CL	NF
LF	2.87	6.18	2.87	1.51	0.30	5.30
SI	6.18	13.34	6.18	3.25	0.65	11.39
LI	2.87	6.18	2.87	1.51	0.30	5.28
TF	1.51	3.25	1.51	0.79	0.16	2.78
CL	0.30	0.65	0.30	0.16	0.03	0.56
NF	5.30	11.89	5.28	2.78	0.56	9.72

2. GINGERICH'S METHOD

A) TALLY MATRIX

	LF	SI	LI	TF	CL	NF
LF	0	8	0	2	0	8
SI	10	0	5	2	0	11
LI	0	8	0	1	2	4
TF	1	2	3	0	0	4
CL	1	1	0	0	0	0
NF	6	9	7	5	0	0

B) PROBABILITY MATRIX

	LF	SI	LI	TF	CL	NF
LF	0	.44	0	.11	0	.44
SI	.36	0	.18	.07	0	.39
LI	0	.53	0	.07	.13	.27
TF	.1	.2	.3	0	0	.4
CL	.5	.5	0	0	0	0
NF	.22	.33	.26	.19	0	0

C) PREDICTED RANDOM MATRIX

	LF	SI	LI	TF	CL	NF
LF	0	.34	.18	.12	.02	.33
SI	.25	0	.31	.14	.03	.37
LI	21	.33	0	.12	.02	.31
TF	.20	.31	.16	0	.02	.30
CL	.18	.28	.15	.10	0	.27
NF	.24	.38	.20	.14	.03	0

Total of 100 (+1) transitions

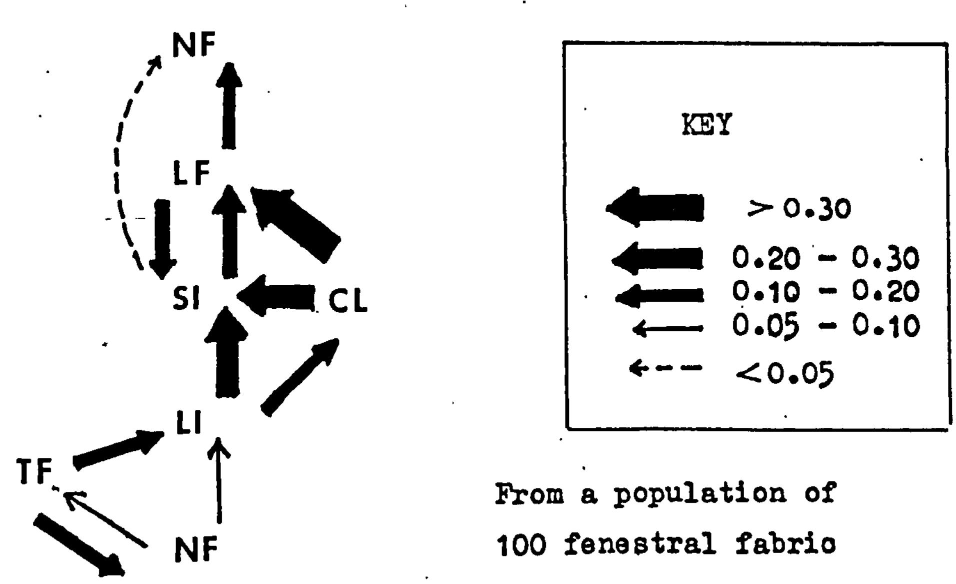
Fig 45 Matrix showing the residual transition probability after the independant random transition probability matrix has been taken from the observed transition probability matrix. This method does not include multistorey events. (Gingerich's method.)

Abbreviations according to Fig. 44.

	LF	SI	LI	TF	CL	NF
LF	0	+0.10	−0.18	−0.01	−0.02	+0.11
SI	+0.11	0	−0.03	−0.07	−0.03	+0.02
LI	−0.21	+0.23	0	−0.05	+0.11	−0.04
TF	−0.10	−0.11	+0.14	0	−0.02	+0.10
CL	+0.32	+0.22	−0.15	−0.10	0	−0.27
NF	−0.02	−0.05	+0.06	+0.05	−0.03	0

(see fig 44a for probability and tally matrices)

Fenestral fabric relationship diagram from the above matrix showing the probabilities of upward transitions from each fenestral fabric, after allowance for random transitions (Gingerich's method).



From a population of 100 fenestral fabric association transitions, commencing with NF.

dominated by tubular fenestrae, comparing their vertical association to Shark Bay analogues. In contrast, the Ordovician fabrics of Grover and Read (1978) lack cyclicity and are commonly intimately associated together.

Below are reviewed the probable palaeoenvironments of these Lower Carboniferous fenestral fabrics in comparison to Recent analogues.

6.4.5. Fenestra palaeoenvironments

6.4.5.1. TUBULAR FENESTRAE occur associated with root systems of salt-tolerant (halophyte) land plants inhabiting the upper intertidal to supratidal zones in Shark Bay. These fenestrae (Logan, 1974, p236-237) range from 1 to 5mm diameter and are vertically interconnected. Push-apart formation of these fenestrae is indicated by tighter grain-packing around the open tubules.

In contrast, Grover and Read (1978) attribute their 0.1 to 2mm diameter branching tubules to infauna, likening them to the intertidal burrows of Ginsburg and Hardie (1975) and stressing the common association of vertical burrows within intertidal sediments (e.g. Walker, 1972). Logan (1974 p.228) recognised intense boring activity in lithified sustrates from a species of small bivalve. The lack of any shelly remains within these Asbian tubules suggests their formation by a similar organism is unlikely. Garrett (1977) noted that tubular fenestrae of comparable size form in Recent intertidal flats by burrowing activities of both polychaetes and nematodes. Fine peloidal internal sediments occur in many fenestral fabrics, but are more common in larger non tubular forms indicating that if they are of faecal origin (see next section) the producing organisms prefered to inhabit/formed non-tubular fenestrae.

Tubular fenestrae rarely occur as a second generation fenestral fabric, overprinting pre-existing fenestrae (e.g.s/m 289; 1498 where they underlie discontinuity surfaces (see section 11. 1) indicating reactivation associated with transgression following a period of emergence, i.e. in these instances, the fenestral 'cycle' is reset(?) during the transgression.

In specimen 1472 (Sychtyn Member) prominent vertical tubules occur that have side branches and allow admixing of overlying clays down the open tubule. Associated with ped-structure, (see section 6. 4. 13) they indicate that some tubular fenestrae have probable rootlet origins (see section 11.3.2.2).

In conclusion, the common occurrence of tubular fenestrae in MA.1 lithologies lacking other fenestral features and their (?non random) position at the base of fenestrated sequences (figs.44 & 45) suggests that most have a shallow subtidal to low intertidal infaunal burrow origin.

6.4.5.2. LARGE IRREGULAR FENESTRAE occur towards fenestra unit bases according to the fabric relationship diagrams (figs.44&45). They are normally sublaminoid, and may be directly analogous to the fenestral fabrics associated with colloform algal mats from the subtidal zone of Shark Bay (Logan, Hoffman & Gebelein, 1974), that are formed through a combination of growth (of algal mat) expansion, binding, cementation and oxidation. These fenestral fabrics are, however, normally associated with unlaminated or poorly laminated MA.1.1. However, lamination of algal mat associated sediments is a result of episodic sediment influx (Logan, Hoffman & Gebelein, 1974, p.152) and variations in type and density of organic material. In a steady state situation (especially a subtidal setting) unlaminated algal-induced sedimentation may ensue.

6.4.5.3. SMALL TO MEDIUM IRREGULAR FENESTRAE are the most common fenestral fabrics present in MA.1 suites. Similar fabrics are well documented throughout the geologic record (Deelman, 1972). Shinn (1968) formed them though periodic desiccation of homogenous lime muds Logan, Hoffman & Gebelein (1974) recognised their association with pustular algal mats from middle and upper intertidal zones at Shark Bay, and Kendall and Skipwith (1968) from similar algal fabrics in the Arabian Gulf. As with the Shark Bay fabrics, these Asbian fenestrae are associated with unlaminated grain-supported sediments. The presence of stacked concave-up fenestra-sets is indicative of a sediment surface microrelief due to the formation of 'pustular' or 'cinder' mat fabrics. This fenestral fabric, occurring commonly as thin densely fenestrated units, is interpreted as an intertidal fabric.

6.4.5.4. LAMINOID FENESTRAE are associated with smooth algal mats in the lower intertidal zone at Shark Bay. This analogy would place them below the small to medium irregular forms if a consistent shoaling model is followed. According to figs. 44 and 45, they are often intimately associated with the irregular fenestral fabrics (fig.43). The occasional development of stromatolitic relief within this fabric indicates formation within the intertidal to subtidal zones,(due to(?) tidal flood and ebb scouring) although stromatolite tops may accrete within the supratidal zone (cf. the 'film mat' (Woods and Brown,1975)). The lack of desiccation cracks associated with the laminoid fenestrae also indicates no prolonged emergence occurred. Generally associated within layered sediments, they suggest a high position on the tidal flat where storm-produced laminites are common (cf. Recent Andros tidal flats).

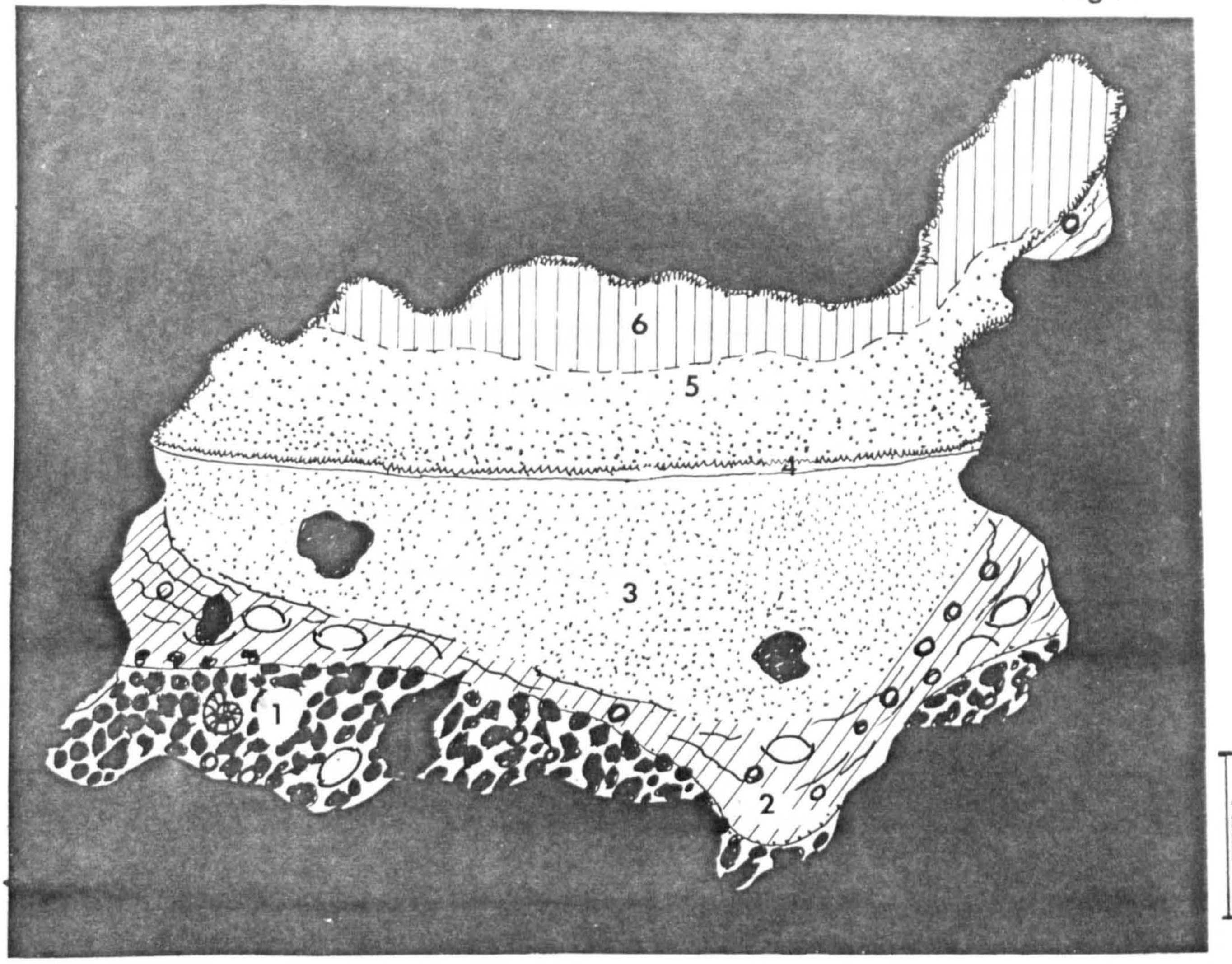
6.4.5.5. IMPLICATIONS The stratigraphic sequence (see fig 43) of fenestral fabrics suggests that the fenestral fabrics reflect sequential & lateral variations in palaeoenvironments, or that the fabric types were random within palaeoenvironments. The fabric relationship diagrams show a vertical fabric cyclicity and association comparable to Recent peritidal environments, when considered within a progradational model: from tubular and large irregular fenestral fabrics of shallow subtidal to low intertidal origin to medium and small irregular fenestrae of intertidal origin, and laminoid fabrics, associated with laminite sediments of high intertidal/supratidal environments. Further data may elucidate the validity of this fabric association model to a higher degree of confidence.

6.4.6. Geopetal (Internal) Sediments

Within penecontemporary open-space structures (fenestral fabrics, umbrella voids and fossil chambers) a varied selection of internal sediments are present that infill or partly infill these pore spaces in a sequential manner (fig. 46).

6.4.6.1 GRAIN-SUPPORTED SEDIMENT. In some of the larger pore-spaces a basal layer of grain-supported sediment occurs which in exceptional cases completely infills the pore (e.g. s/m 2528). Rarely it may overlie mud-supported internal sediment (e.g. s/m 1477). The sediment may be derived (depending upon its character) from the surrounding sediment or from an ?overlying source. It is usually a peloidal grainstone, with well sorted clasts, and occasional larger clasts floating within, that have spalled off the cavity roof. The peloids are rounded to angular, and are mostly 20 um to 50 um diameter. S/m 289 and 2528 contain bioclastic grainstones that have infiltrated up to tens of centimetres from overlying transgressive units (MA.3)

6.4.6.2 LIME MUD SUPPORTED SEDIMENT This internal sediment also lines cavity bases and has internal lamination indicating periodic



Fenestral cavity showing observed sequence of internal sedimentation (composite and diagrammatic)

KEY

1	Grain-supported 'pellet silt'.
2	Matrix-supported wackestone, with lamination and bioclastic debris, including ostracods.
3	fine crystal silts .
4	Isopachous rim of acicular carbonate cement.
5	Crystal silts, grading coarser.
6	Spar druse.

influx. Nests of whole ostracods occur in a few cavities, even in ostracod-deficient sediments (Plate 6 fig. E). Similar occurrences have been reported from stromatactis cavities in allochthonous blocks of Asbian mudmounds in Ecton Limestone, Swainsley, Derbyshire (Aitkenhead and Chisholm, 1979)†. Garett (1977) described infaunal ostracods in 'pond' environments on Andros tidal flats. Cavity systems may have provided an ideal 'retreat' for such infauna in the ancient. It is usually this infill that is burrowed by succeeding generations of tubular fenestrae. When present, this sediment grades upwards into crystal silts.

6.4.6.3. CRYSTAL SILTS. Overlying mud-supported sediments, are variable quantities of crystal silts. Fenestral cavities normally possess this geopetal filling only. The crystal silts grade in size from micrite grade material at their base to equant mozaic microspar/pseudospar (c. 20μm) at the top. Mostly there is a sharp top to this silt drape, above which spar druse infills the cavity.

Below vertical elements of cavities this silt may be stacked in small piles. It also commonly clings to steep (even vertical) cavity walls indicating a very high coefficient of friction, probably enhanced by contemporaneous cementation.

6.4.6.4. Significance of internal sediments

Dunham (1969a) recognised the mechanically deposited nature of crystal silts overlying early cement fabrics within cavities, and argued for their deposition and transport by flushing waters in a vadose environment. The silts are most prolific in fenestral fabrics that have a high degree of vertical interconnection. Tubular fenestrae provide channels down which silt-laden waters could readily percolate. The presence of isopachous cements (see next section) interlaminated with crystal silts suggests that

† Pal. Ass. Carboniferous Studies Group, Field Meeting, Easter, 1979 Fieldguide.

periods of non-deposition occurred in a phreatic environment, when pore-waters, with a high PCO_2 (Matthews, 1971) (and probably associated with a low Eh) readily precipitated cements. The upward increase in silt size within cavities (often only marked towards the top of the silt pile) remains a speculative problem. Dunham (1969a) derived the silts from erosion of host sediment and cement in the vadose zone. As a corollary to this model it is significant to note that as lithification continues, fewer pathways are available for silt transport, and, in general, cement crystals in the larger cavities, growing centripetally have an increasing size. These larger crystals are then more susceptible to erosion as the waters flush down their decreased gape. Dunham (1969a) realised that phreatic currents are generally too slow to erode and transport this silt-grade material.

Grover and Read (1978) noted floors of fenestrae lined with 'Pellet Silt' (Dunham 1969a) of heterogenous origin. The occurrence of grain-supported silt in the Tynant Formation is limited to units underlying MA.2 or MA.3. Their deposition is envisaged as an early event, subsequent to the fenestra formation, flushed down up to tens of centimetres into an open, interconnected cavity system. Erosion† associated with the transgressive event laid-bare cavity systems which provided sediment traps. This pellet silt often bears little lithological similarity to adjacent sediments (Dunham, 1969a, p.151) although in MA.1. lithologies ~ 30µm peloids are common indicating that erosion and transport may have had a sorting effect, with only the larger particles being trapped in some cavities.

EARLY CEMENTS AND REPLACEMENTS

Penecontemporaneous dolomite, vadose and phreatic carbonate cements are recognised associated with MA.1.

† and intertidal dissolution (II.1.)

6.4.7. Early lithification of Lime Mud

The presence of intraformational conglomerates and 'sutured discontinuity surfaces' (see section 11. 1) indicate that some lime muds and silts were at least partly lithified at a penecontemporaneous stage by microcrystalline intergranular cement. Further evidence is shown for early lithification in s/m 2109, in which cryptalgal laminites are disrupted into angular, centimetre-sized flat clasts, that ride over each other, in a zone descending into the sediment at 45° to the lamination. This zone of disruption is about 10cm deep, and many clasts show that they originally interlocked (Fig. 38). Wackestone infills the matrix to these intraclasts. Assereto and Kendall (1977) have described similar structures associated with similar sediment suites as 'peritidal tepees', formed through lithification-expansion, enhanced by desiccation. Such tepees normally have a polygonal structure in three dimensions, but this has not been recognised in this example (cf. Scherk,1975)

Logan (1974 p213) described cryptocrystalline aragonite cementation on intertidal veneers that produced cohesive sediments whilst immersed in 'tidal brines' (39‰ Cl), but which hardened readily on drying and through this desiccation process,with erosion produced intraclastic conglomerates. Pratt (1979) also described high magnesian calcite micritic cements in intertidal environments, associated with cryptalgal fabrics.

The lack of convincing desiccation polygons throughout Asbian MA.1. microfacies implies that these sediments may have been lithified by microcrystalline cementation from interstitial 'tidal brine' waters.

The following suite of early pore-filling cements recognised in MA.1. lithologies supports this interpretation, although vadose and meteoric phreatic waters may also have been in-part responsible.

6.4.8. Penecontemporary Dolomite

This occurs within MA.1.2b in the Tynant Limestone at Tynant (G.R.SJ21954570), and within intraclasts of MA.1.1 in the Tynant Limestone at G.R. SJ23424948. Other more problematic examples are documented. (Plate 7, fig. D)

A. At the Tynant locality the dolomite has a localised distribution, occurring as millimetre to centimetre thick partly dolomitic ochrous layers within finely laminated calcisphere-peloid grainstone-packstones (fig 47).

These ochrous layers are rich in 2µm to 10µm (Plate 7, fig. B) sub-rhombohedral dolomite (10- 80% by volume), highlighted in alizarin red-S stained acetate peels (Plate 7, fig. F). The basal layer-boundaries are diffuse as the relative proportions of calcite to dolomite increase downwards into non-dolomitic layers (cf Bose, (1979 p685)). The dolomite-rich layers are persistent over centimetres to metres of outcrop. Where dolomitisation is incomplete, clast boundaries are enhanced. Bioturbation has piped the layers into each other. Dolomite-rich intraclasts† occur within the dolomicrospar deficient layers (and vica versa to a lesser extent) as incontrovertible evidence for early dolomitisation (e.g.Germann,1969)

Some grains are only partly replaced by dolomite, leaving calcitic cores. The dolomitisation has mainly affected the matrix.

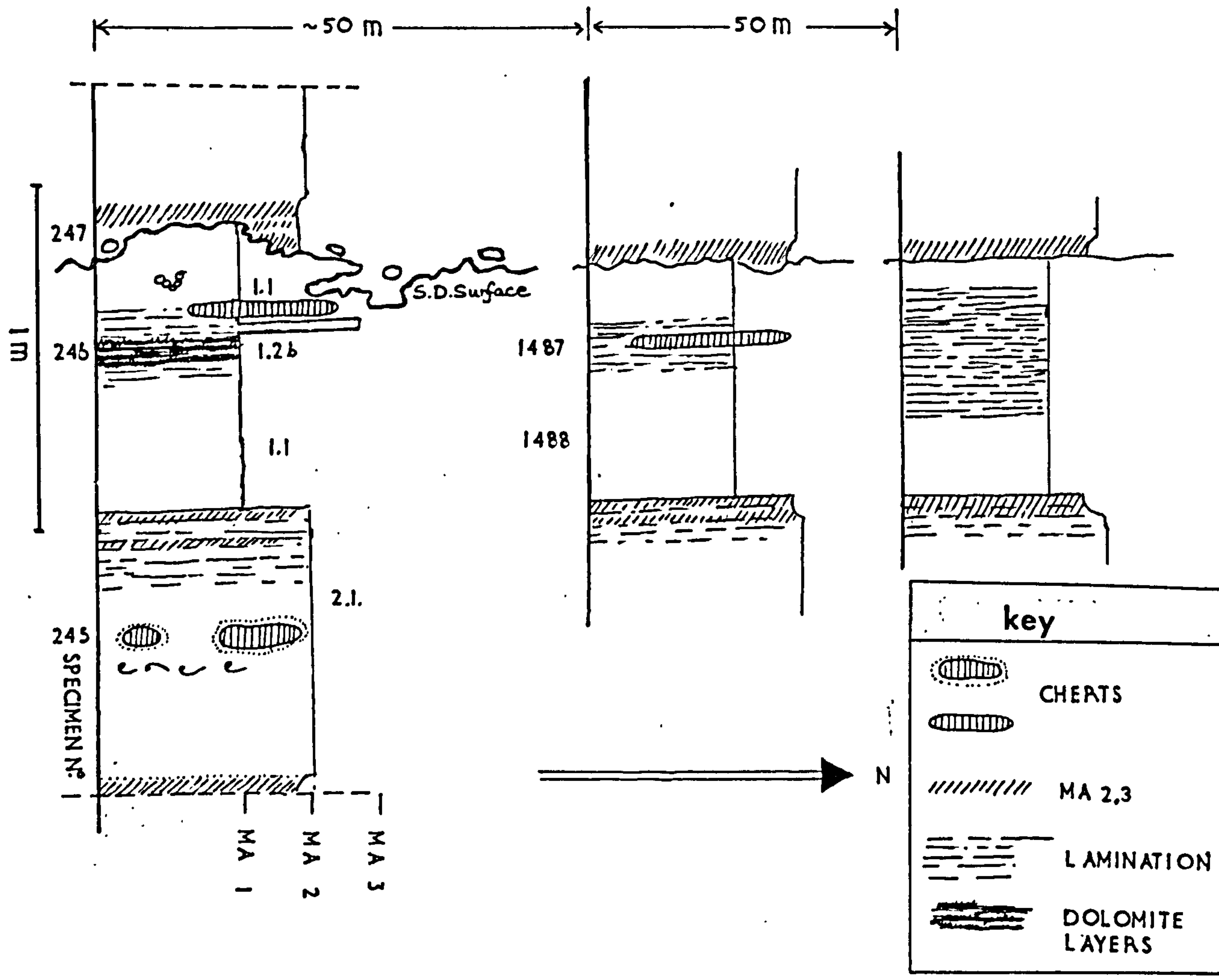
Fenestral cavities and grainstone laminae that are associated with the dolomitic layers have poorly developed fringes of clear 20µm to 100µm dolomite rhombs that may be an early dolomite cement (Plate 7, figs.C&E). The central areas of the fenestrae are clear non-ferroan drusy calcite. Rare dolomite crystals 'floating' within the spar druse may be due to a '3D' effect with respect to the section position, (see Plate 7, fig. E). In more dolomitic laminae non-micritic matrix is now 10µm to 50µm dolomite(?)primary cement.

† Plate 7, fig. A

Fig 47

STRATIGRAPHIC SETTING OF PENECONTEMPORANEOUS DOLOMITE

TYNANT FORMATION, TYNANT.



G. R. SJ 21954570

PLATE 7 Penecontemporary Dolomite

A Penecontemporary dolomite lamination (dark) and calcitic lamination (light) in laminated calcisphere-peloid packstone-grainstone. Note dolomitic intraclasts. s/m 246, Tynant Formation, G. R. SJ 21984572.

B S.E.M. photomicrograph of dolomitic lamina with fenestral 'cavity' infilled by coarse calcite spar. Calcisphere towards bottom right is still vaguely visible. S/m 246, Tynant Formation, 10% HCl etch for 60 sec. .

C S.E.M. photomicrograph of early dolomitic cement lining roof of fenestral cavity. Rhombs are well formed, but elongate pits (C.lum) suggest inclusions. 10% HCl etch for 60 sec. . s/m 246, Tynant Formation.

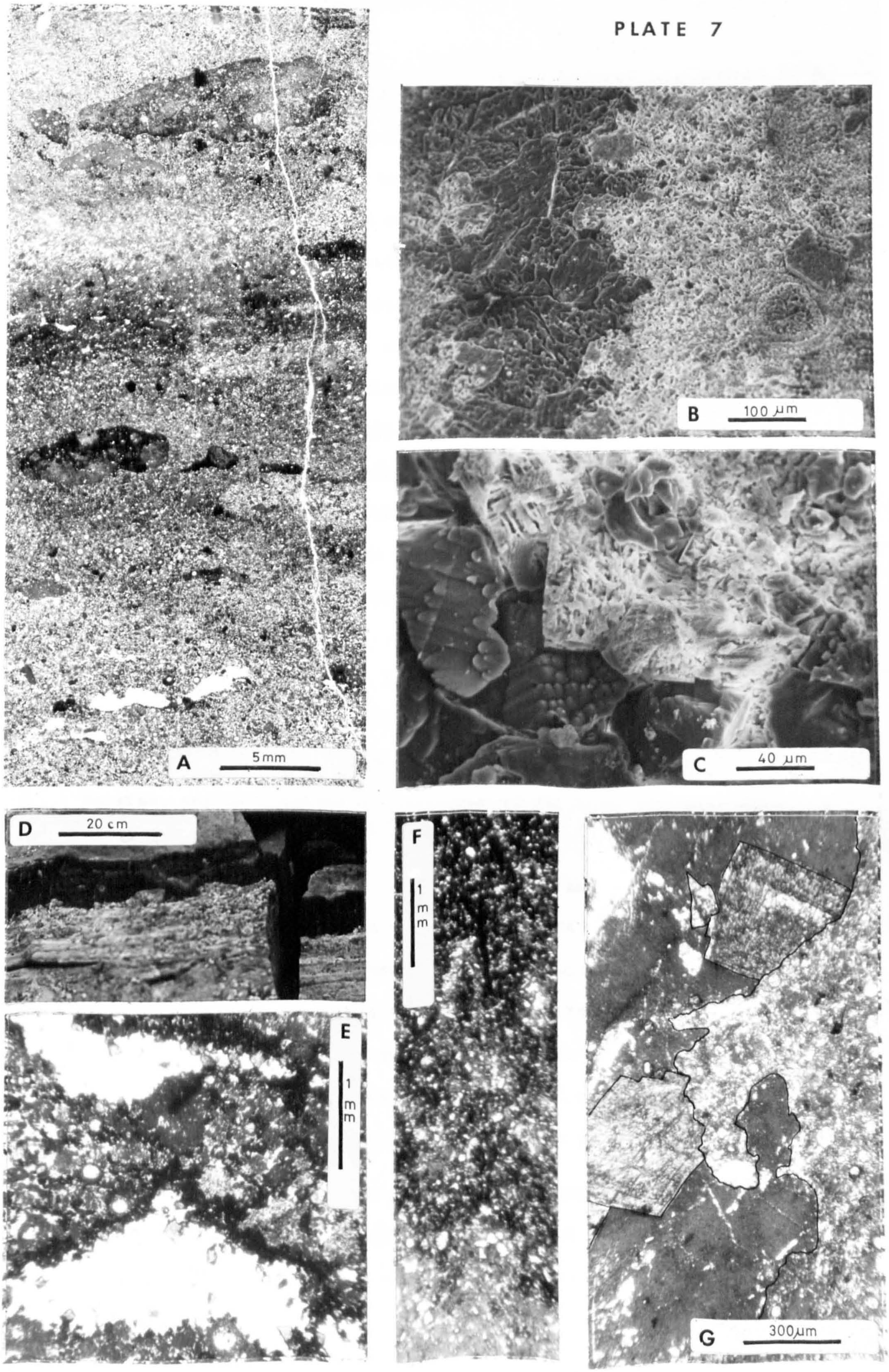
D Penecontemporary dolomite lamination visible (field photo) as ochre (pale) laminae, underlying a tabular chert (black) nodule Tynant Formation, s/m 246 locality.

E Fenestral fabric within partly dolomitised layer. Note greater replacive dolomitisation of sediments above cavity roofs, and dolomite cement apparently floating in fenestra (cf.fig. C) s/m 246, Tynant Formation.

F Grading contact between dolomitic and non-dolomitic layers, s/m 246, Tynant Formation.

G Large zoned dolomite rhombs in intraclast (outlined). Dolomite is eroded at intraclast margin. s/m 192, Tynant Formation SJ23424947

PLATE 7



B. In specimen 192 large intraclasts of MA.1.1 occur within MA.3. (chapter 7). These intraclasts are partly dolomitised by scattered replacive 100μm to 500μm zoned dolomite rhombs. The rhombs have been fractured and eroded where they occur on intraclast margins. This is interpreted as evidence of early dolomitisation prior to erosion and redeposition as intraclastic material (Plate 7, fig. G). The rhombs are well-formed, zoned, have curved faces and are rarely in 1mm aggregations. Occasional clasts of detrital dolomite attest to its redeposition.

C. Specimen 1511 (a scree block from Tynant) of typical Tynant Formation MA1.1. provides some tantalising evidence of early dolomitisation. Selective dolomitisation is well shown associated with a laminated calcrete crust (with a prominent alveolar texture (see section 11.3.1.2.)). The dolomite crystals (5 to 200μm) preferentially replace the lime mud and bioclasts, but not the calcrete crust. Partly dolomitised intraclasts suggest a penecontemporary formation. Fenestrae have intensely dolomitised margins, and nucleations of disseminated dolomite may represent burrow infills.

D. Discussion In the Arabian Gulf (Illing et al., 1965), Andros Island (Shinn et.al., 1965) and Bonaire (Deffeyes et al., 1965) 'dolomitic' crusts form contemporaneously and penecontemporaneously on supratidal flats. Behrans and Land (1970) described penecontemporary subtidal 'dolomite' in Baffin Bay, Texas.

In Bonaire and Andros the 'dolomites' occur as lime mud undergoing progressive dolomitisation by the growth of 5μm protodolomite crystallites at the expense of aragonite.

Dolomitisation on Andros occurs supratidally on tidal-channel levee crests and back-slopes associated with pelleted muds, stromatolitic mats and desiccation cracks, and is most concentrated on surface-hardened crusts. The protodolomite replaces aragonite muds and acts as a cement along with calcite and aragonite. Sedimentation occurs during rare tidal and storm flooding. The dolomitic crust is between 1cm and 5cm thick, and is readily broken up into rip-up clasts (Shinn et al., 1965). The Arabian Gulf penecontemporaneous dolomites are forming in supratidal sabkhas, as 'true' ordered dolomite (Illing et al.1965), replacing aragonite, and decreasing in extent of dolomitisation with depth into the supratidal sediment wedge.

The lack of sabkha sediments associated with these Asbian dolomites preclude their formation in an Arabian Gulf situation. Deffeyes et al. (1965) emphasised (p.87) the need for increased Mg/Ca ratios with respect to normal sea water to form protodolomite. Precipitation of calcium sulphates increases the Mg/Ca ratio in the Bonaire example, allowing evaporative pumping (Hsu e Siegenthaler,1969;1971) to provide a seepage-transport mechanism for replacive dolomitisation.

Bose (1979) recognised that compositional variation and permeability controlled selective dolomitisation, thereby accounting for calcitic intraclasts and clasts only partly dolomitised on their outer margins. He (op. cit. p.688) also recognised that 'fossil' counterparts of penecontemporary 'dolomites' have a coarser crystal size, "explained by aggrading recrystallisation of the original dolomicrite", and that 'dolomite' may be directly precipitated as void-filling cement as a secondary process to replacement. Graf and Goldsmith (1956) suggested that many ancient dolomites may have gone through a slow reordering phase from original protodolomites.

Protodolomite can only form if the Mg/Ca ratio exceeds 5-10 : 1 in hypersaline environments, whereas at reduced salinites it is able to nucleate at Mg/Ca ratios approaching 1 : 1 (Folk & Land 1975). Gebelein et al (1979) provided a model for the formation of protodolomite in a mixing-zone environment between fresh-water phreatic lenses and the marine phreatic zone on Andros tidal flats. Deffeyes et al. (1965) suggested that refluxing dense brines (with a high Mg/Ca ratio) may be responsible for more pervasive 'secondary' dolomitisation, where overlain spatially by an emergent, evaporative terrain. Zenger (1972) however pointed out the pit-falls of applying a supratidal penecontemporary model to many ancient dolomites.

The Tynant locality dolomite contains many petrographic features diagnostic of penecontemporary formation. Its position (see fig. 47) towards the top of a shoaling cycle, intimately associated with cryptalgal fabrics substantiates this claim. Lateral extent is estimated in hundreds of metres rather than kilometres, showing that it is a rare localised event.

Reports of other ancient laminated dolomite calcite sediments include the Ordovician New Market Limestone (Matter 1967) Mississippian carbonates by Schenk (1967). Triassic loferites by Fischer (1975) and the Lower Devonian Manlius formation (Laporte, 1975).

At the tops of other Tynant Formation cycles, dolomitisation occurs on a more pervasive scale, with 20 to 200μm dolomite rhombs scattered to dominant. There is no significant evidence of a penecontemporary origin for these, although association with fenestrated MA.1 erosive surfaces and similarty with s/m 192 type dolomitisation suggest that an early, diagenetic origin cannot be ruled out (see Chart A).

The lack of associated evaporitic minerals and evidence of their formation suggests that hypersalinities did not exist. Bourrouilh (1978) noted that tidal ponds of Andros are alkali hypersodic (through dissolution of carbonate and organic decomposition (Gebelein 1978)) and therefore not capable of precipitating gypsum. Although MA.1.2b resembles recent storm deposits on supratidal flats and levees (Shinn et al., 1969; see section 6. 6. 3.) there is no evidence of desiccation, however evaporative pumping may keep sediments damp, whilst nucleating (proto-) dolomite. Hardie and Ginsburg (1977 p.59) noted that desiccation cracks may be absent from supratidal sediments. If the precipitation rate was high enough to produce fresh water lenses (the presence of karstic erosion features and lack of deep calcrete profiles suggests that this was so throughout much of the Asbian) on tidal-supratidal flats then dolomitisation may have occurred in a mixing zone environment (Dorag-type model of Badiozamani (1973)) as suggested by Havard and Oldershaw (1976) for their Devonian dolomitic facies. This is further supported by s/m 1511 in which incipient dolomitisation may have occurred beneath an actively forming calcrete crust. Calcrete crusts formed in supratidal environments associated with plant colonisation. Meteoric intestitial waters are obligatory in calcrete environments, but in this example proximity of marine phreatic waters is likely, therefore providing an ideal mixing zone situation.

6.4.9. Isopachous Acicular, Bladed and Equant Cements

(Terminology following Havard and Oldershaw (1976)).

This cement type occurs within both MA.1 and MA.3 sediments, lining primary and penecontemporary cavities. It is often associated with micritic cements and rarely with vadose dripstone fabrics. When internal sediments are present it may underlie, overlie or be interbedded with them (e.g. s/m 1325, Sychtyn Member).

The crystals are clear to dusty non-ferroan calcite, and normally about 40μm long by 20-30μm wide, (max. 500μm fringe in s/m 1215, Whitehaven, Sychtyn Member), all similarly orientated (centripetal to the cavity), some equant with bladed terminations, and others retaining acicular forms marked by either micritic cement or drusy spar (Plate 8, fig. A). Their distribution is very localised (e.g. in s/m 1456, where only a few irregular fenestrae show the rim upon a thin veneer of geopetal silt.). These bladed cement fringes occasionally have indistinct thickening on cavity roofs. This cement fabric is morphologically similar to many Recent beach-rock high-magnesian calcite cements (Tietz & Muller, 1971)

The problem rests on whether or not the crystal/cavity interface can retain the crystallographic form of the original cement. In s/m 092 (MA.3.3. microfacies) fine acicular terminations are preserved in a micrite cement coating, whereas in s/m 1590 bladed terminations are similarly preserved. Bladed terminations mostly occur when this cement is coated with crystal silts† but retain acicular form when overlain by coarse spar druse. This therefore suggests that the primary crystal morphologies were acicular and that subsequent recrystallisation converts the crystals to bladed morphologies. Some of the bladed fringes may, however, be primary.

Fine acicular terminations suggest an original high-magnesian calcite or aragonitic mineralogy. Isopachous fringes suggest formation from phreatic solutions. Interposition with vadose internal sediments indicates sequential variation between vadose and phreatic conditions, presumably due to intermittent rise in the water table. Badiozamani _et al_. (1977) showed that under laboratory 'phreatic' conditions Mg++ ions favour precipitation of fine acicular calcites, whereas euhedral bladed to equant calcites were formed in Mg++ deficient pore waters. Hanor (1978) showed experimentally that 30µm low Mg calcite cement rinds were formed from mixed meteoric-

† Plate 8, fig. B

PLATE 8 Early Cements.

A Acicular (non-ferroan) cement rim to fenestral cavity within MA.1.1. s/m 2564. Tynant Formation, Froncysyllte G. R. SJ 26984199

B Fringes of equant and bladed clear calcite interbedded with internal sediments. S/m 1325. Sychtyn Member, Craig Sychtyn, G. R. SJ 233261.

C Early (ferroan) acicular cement fringe, overlying an isopachous fibrous fringe, s/m 2563, Tynant Formation G.R. SJ 26984199

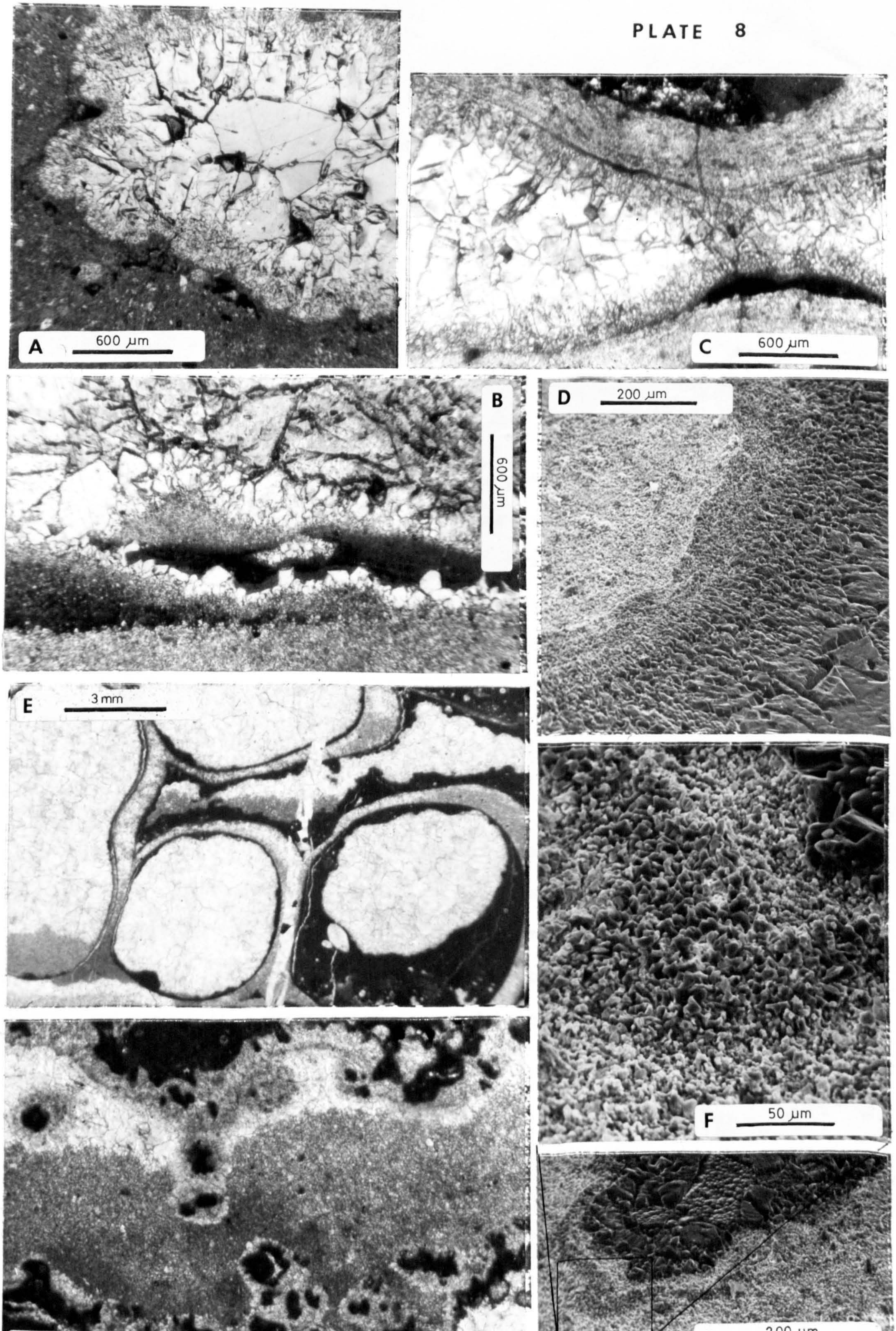
D S.E.M. photomicrograph of acicular cement of fig. C.

E Irregular coating of dense (pale) micrite cement within and external to a gastropod. Cementation predates geopetal silts. Note incorporation of ostracod and thickening of cement on cavity bases. Tynant Formation, s/m 027 Tynant G. R. SJ 22064533

F & G S.E.M. photomicrographs of isopachous cements to fenestral margins, comprising: an inner (dark) fringe of silt grade cement (?from an acicular precursor); an outer (pale) fringe of structureless micrite (equivalent petrographically to micrite cement of E). S/m 1498, G.R. SJ 22014552

H Microspeleothemic cements (microstalactites and microstalagmites) within fenestral cavity. Note microspeleothem "drowned" by geopetal silts, and dark (clay replacement) centres to the microspeleothems s/m 1590. Froncysyllte G.R. SJ 26984199 Tynant Formation.

PLATE 8



marine waters by CO_2 degassing. It is probable that some bladed Asbian rinds were precipitated in a meteoric phreatic zone (and hence low Mg calcite is the most likely mineralogy) due to their intimate association with both vadose crystal silts and dripstone cements. Conversely the acicular fringes may be relicts of marine beach-rock type cementation(see section 7.3.5.2).

At G.R.SJ26984199 (Sychtyn Member s/m 2563), an early ferroan isopachous acicular rim supersedes fibrous curtain cements lining fenestrae, as 200um fringes contiguous with later non-ferroan coarse equant druse (Plate 8, fig. C). Whether the cement is primary or a replacement of an earlier high magnesian calcite (Richter and Fuchtbauer 1978) is not clear. As a primary (presumably low magnesian) ferroan calcite, it would have to be precipitated from reducing phreatic waters. Association with illitic replacements implies that ferroan pore waters were available. Oldershaw and Scoffin (1967) showed an association of ferroan cements with argillaceous sediments and Richter and Fuchtbauer (1978) suggested clays as a likely iron source.

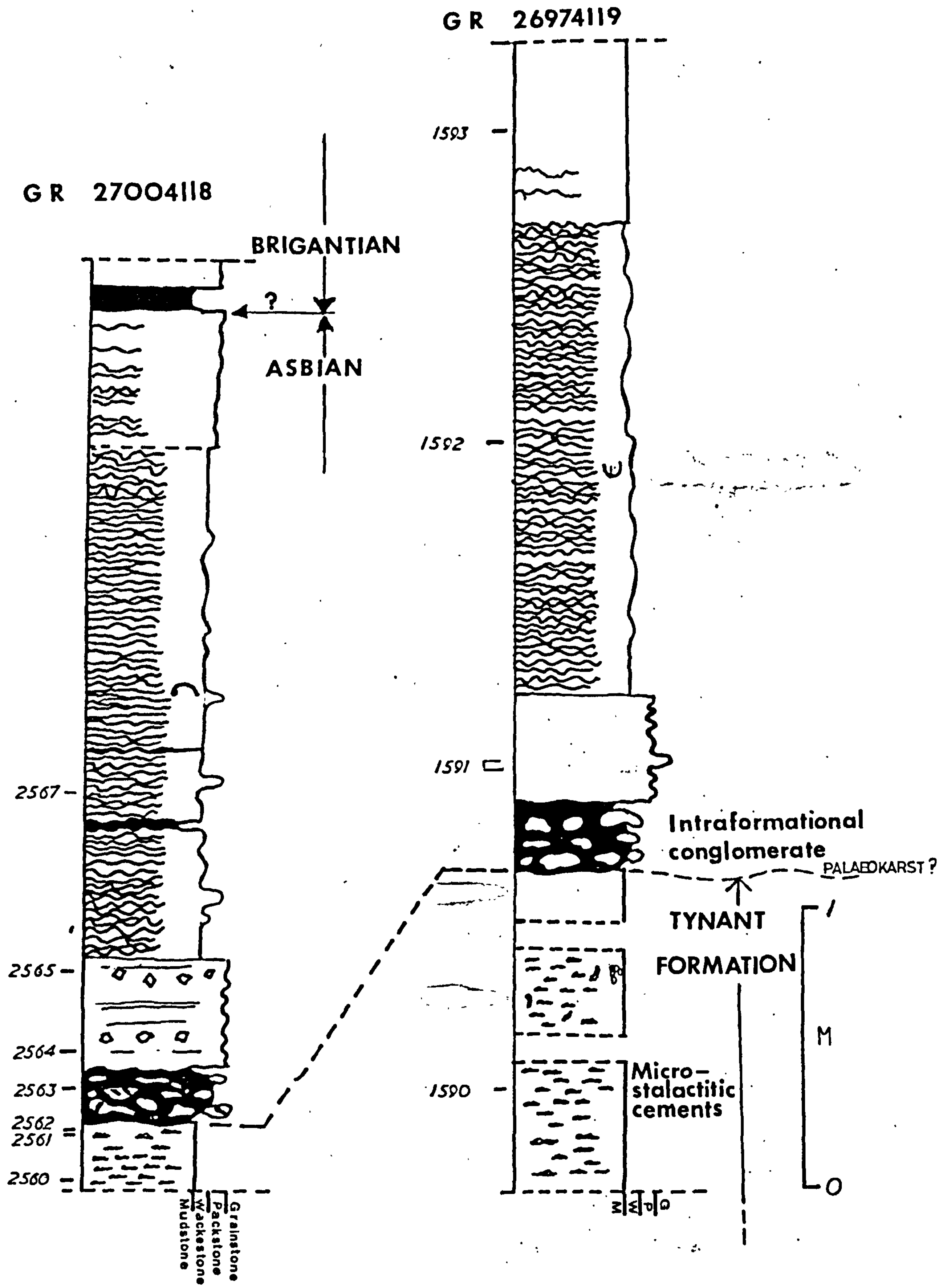
Subaerial weathering of Lower Palaezoic litharenites and argillites may have charged meteoric waters with iron complexes. The association of an overlying karstic surface (fig. 48) indicates that subaerial erosion (and weathering) occurred penecontemporaneously and association with vadose cement fabrics indicates that meteoric waters were present.

6.4.10. Early Micritic Cements.

These cement rinds also occur in MA.3 lithologies, but are rarely developed in association with bladed isopachous cements in MA.1 and MA.4.1 (Plate 8, fig. F & G) lithologies. They are isopachous, but occasionally have an irregular void rim coating,

ASBIAN SUCCESSION AT FRONCYSYLLTE

Fig 48



usually being thicker on cavity bases (Plate 8, fig. E). Micritic cement manifests itself as a dense laminar coating lining penecontemporary cavities and diffusing around clasts into the adjacent sediment. On polished surfaces this appears as a darker rim. The S.E.M. reveals 1 μm to 3μm granular crystallites (Plate 8, fig. F).

Micrite cements (high magnesian calcites) are known from many examples of recent beachrocks (Tiez & Muller, 1971; Moore, 1971); Fursich (1979, p16) has described similar cements from Jurassic submarine hardgrounds. Havard and Oldershaw (1976) noted the presence of micritic rim cements lining pores and interlaminated with both isopachous and microstalactitic cements. They (op. cit., p.62) compared them to calcrete micrites (James, 1972). These micrite cements lack calcrete pigments but resemble the pisolite micrite (section 6. 4.1). This cement fabric is, therefore, probably of 'beach-rock' origin.

6.4.11. Curtains and Microspeleothems.

Microstalactitic, microstalagmitic and curtain (sensu Purser, 1975) cement fabrics occur at two localities[†], in MA.1 microfacies (s/m 1590 and 1102, G.R. SJ26984199 and SJ25202500 respectively).

The microspeleothem cements grew into large irregular fenestral cavities from an original 200μm to 300μm curtain and isopachous fringe of fibrous cement (Plate 9, fig. C). The cement fabric is continuous and interlaminated from curtain to microspeleothem. The cement varies between inclusion dense milky laminae and clearer inclusion deficient laminae. The inclusions are ≤ 1μm, have not been discerned under S.E.M. and impart a vague pseudopleochroism to the host cement. There is no indication of radiaxial twinning.

The curtain and isopachous fringe has ostracod carapaces incorporated within.

The microstalactites have bulbous, rounded basal terminations and are ≤ 1.5mm deep. Microstalagmites are less well developed,

† immediately underlying the Mesothem D5a/D5b boundary of Ramsbottom(1977).

primarily due to their early drowning by thin veneers of internal sediments. They interlaminate with this internal sediment indicating their contemporary formation.

The bulbous nature of the microspeleothems reflects the dominance of curtain cementation, with only localised discontinuous overgrowths of true speleothemic cements within and upon this curtain fabric (Plate 8 fig. H). Microstalactitic and curtain cementation ceased due to eventual drowning by internal crystal silts. (Plate 9, figA)

The presence of inclusions within the curtain and microspeleothemic cements indicates a likely different mineralogy for the primary cement (now presumably low Mg calcite). Both aragonitic (Schneiderman et.al, 1977) and dolomitic inclusions (Meyers & Lohmann, 1978) within cement fabrics have been used to predict precursor mineralogies. High-Mg. calcite may invert to low-Mg calcite, whilst retaining dolomitic inclusions. Due to the unlikely precipitation of aragonite in a vadose (meteoric) situation (this is borne out by the presence of soil profile features associated with these dripstone fabrics (primarily their partial replacement by clays)) high-Mg calcite is a more probable primary mineralogy.

Microspeleothemic cements attest to an 'evolved' vadose environment, primarily by their association with well developed palaeokarstic features. It is significant that the palaeokarst associated with these cements has no calcrete profile (fig. 48), and represents prolonged (?) emergence on the Mesothem D5a, D5b boundary. (Fig. 49)

6.4.12. Palaeopedological phenomena

6.4.12.1. CALCRETE STRUCTURES. At G.R. SJ 22204669 in the Tynant Formation, immediately underlying a cycle boundary, tubular fenestral structures are lined with calcrete micrite (these structures are

CEMENT FABRIC DISTRIBUTION ALONG MESOTHEM D5a/D5b BOUNDARY

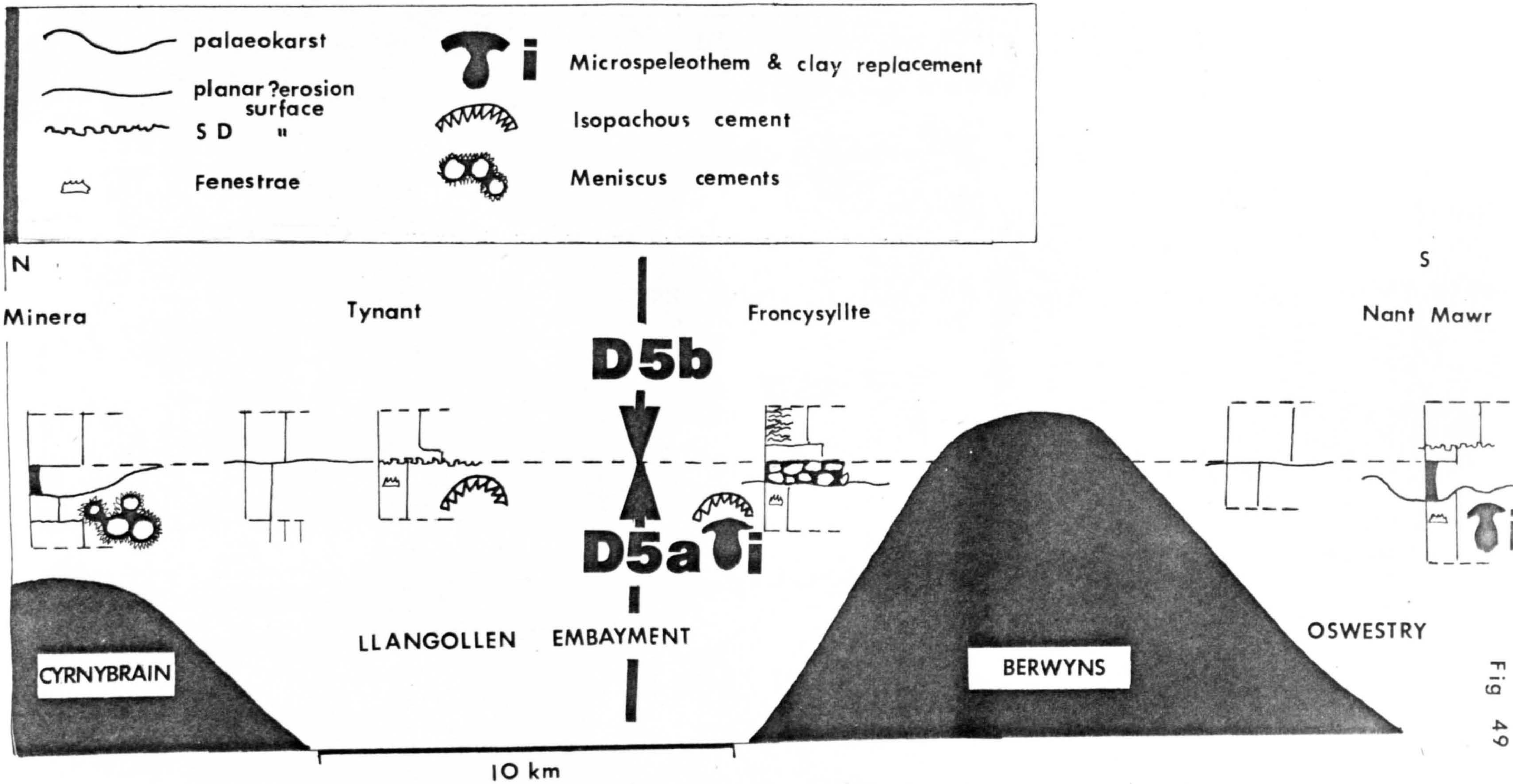


Fig 49

PLATE 9.

A Microstalactite drowned by crystal silt. Note preferential clay replacement within laminae of micro-speleothem and along outer margins. S/m 1590. Tynant Formation, Froncysyllte, G. R. SJ 26984199

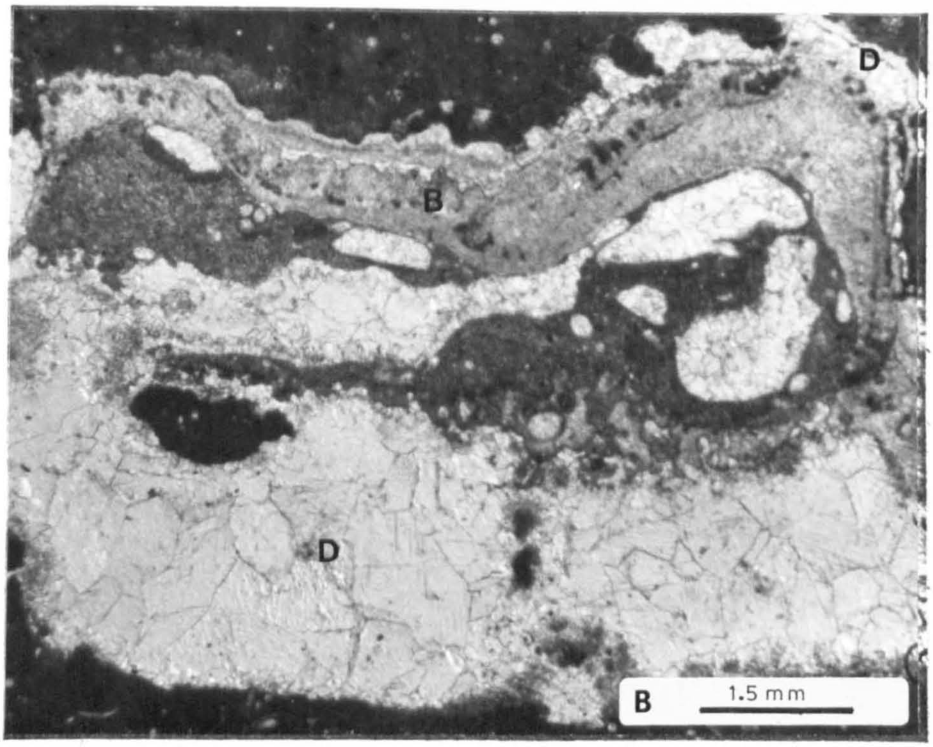
B Curtain cement in fenestrae associated with micro-speleothems (B). Note dendritic clay replacement. Base and margins of fenestrae have coarse clear spar druse (D) S/m. 1590

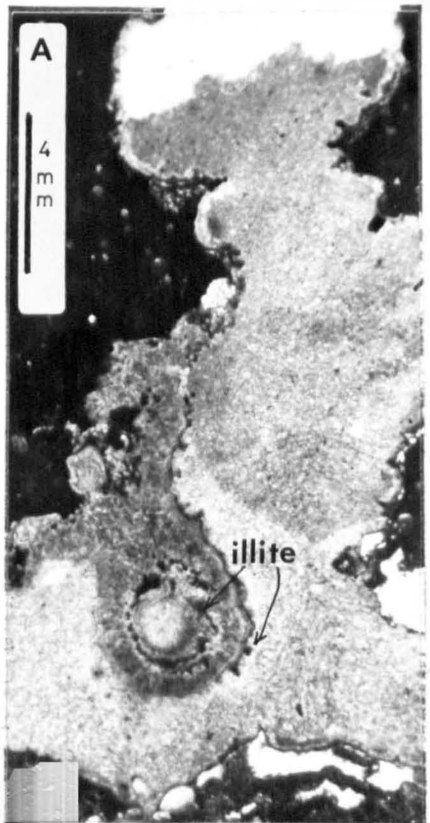
C Fibrous curtain dense with inclusions showing growth lamination s/m 1590

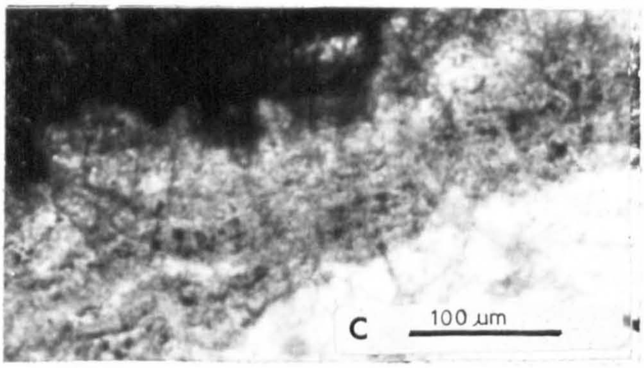
D Pisolitic grainstone. Pale dense micrite coatings to grains, no obvious vadose form. MA.1.2a. S/m 1494. Tynant Formation, Tynant, G. R. SJ 21994561.

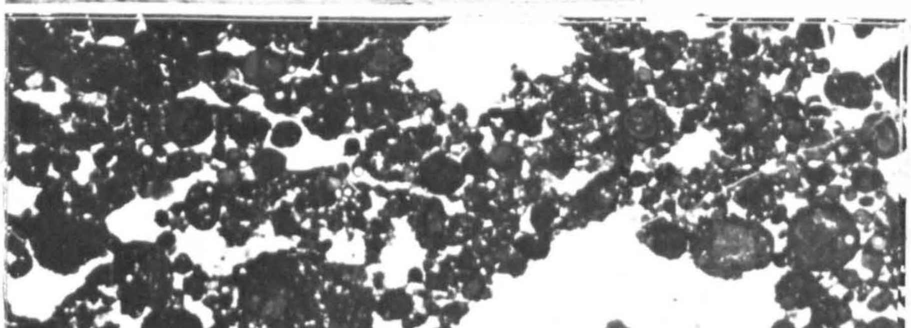
E Deep vertical tubular fenestrae within MA.1.1. Note admixing of clays (black) from above down the open tubule system, and into lateral branches. S/m 1472. Sychtyn Member, Craig y Rhiw, G. R. SJ 236295.

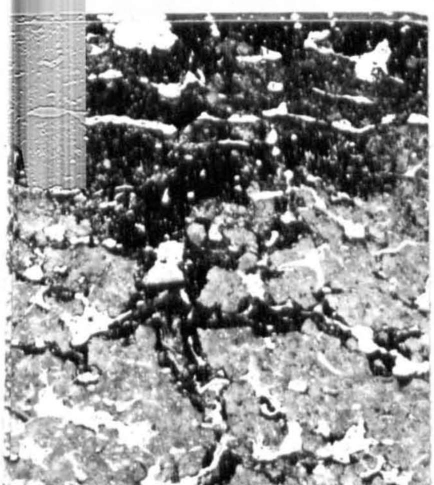
PLATE 9











discussed fully in chapter 11).

In s/m 1511, a laminated calcrete crust is developed within MA.1.1. As this is a loose scree fragment, there is no direct evidence to show Tynant Formation origin.

6.4.12.2. DEVELOPMENT OF PEDS. Fischer (1964 ; 1975) recognised peds of "compound-pelletoid micrite bodies of sediment" that were separated by irregular (fenestral) pore space.

The pseudointraclastic texture produced by packing fenestrae (Plate 6, fig. D) appears as 100μm to 3mm interlocking sediment 'clots' of irregular shape, dependent on the degree of regularity of the packing fenestrae. More laminoid packing fenestrae tend to produce more laterally elongate clots.

Fitzpatrick (1980) described ped structures in modern soils of very similar form (op. cit., p.100), with planes (sensu Brewer, 1964, p.196) intersecting in different directions and with varying degrees of organisation, producing interlocking peds. Such peds vary from 1mm to macropeds 4cm diameter. Sleeman (1963) discussed the origin of planes within soils and attributed them to shrinkage and expansion during wetting and drying, indicating that larger planes only formed under very dry conditions, and that their pattern depends on the kind and uniformity of the soil material and the wetting and drying characteristics.

Whether the clotting of these Asbian sediments is part of a ped-forming process is unclear, but the formation of packing fenestrae suggests a related process. No glaebule (sensu Braithwaite 1975, p.7) structures have been observed in MA.1 lithologies.

6.4.13. Clay Transport within and into MA.1 Lithologies.

At G.R. SJ 23602955 in irregularly fenestrated MA.1.1 clay

PLATE 10

A S.E.M. photomicrograph of irregular clay replacement of carbonate within the microspeleothem / curtain cement fabric of s/m 1590, 1m below Mesothem D5a / D5b boundary, Froncysyllte, G. R. SJ 26984199.

B

C E.D.A.X. X-ray analysis of element distributions

D within the field of fig. A.

E

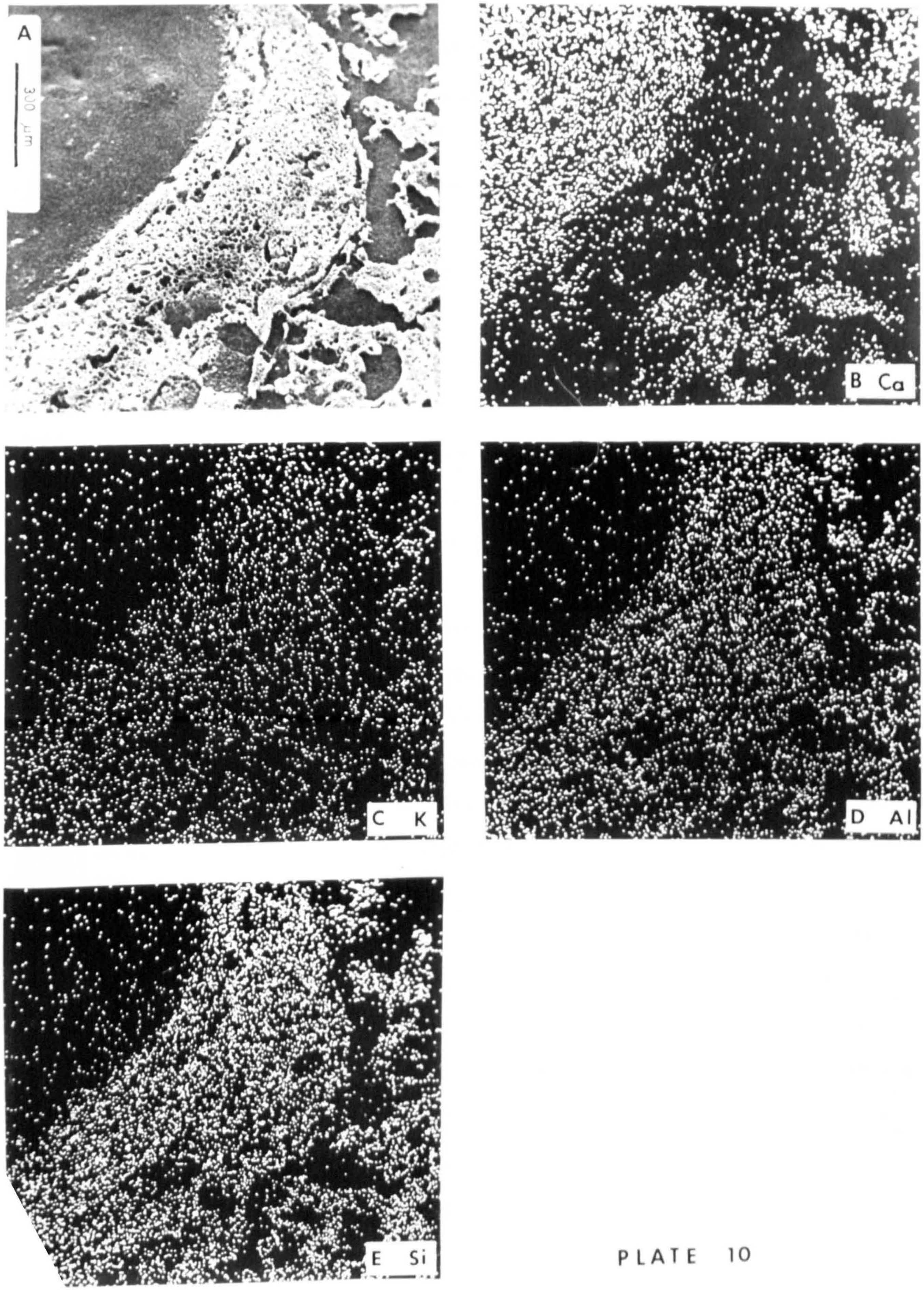


PLATE 10

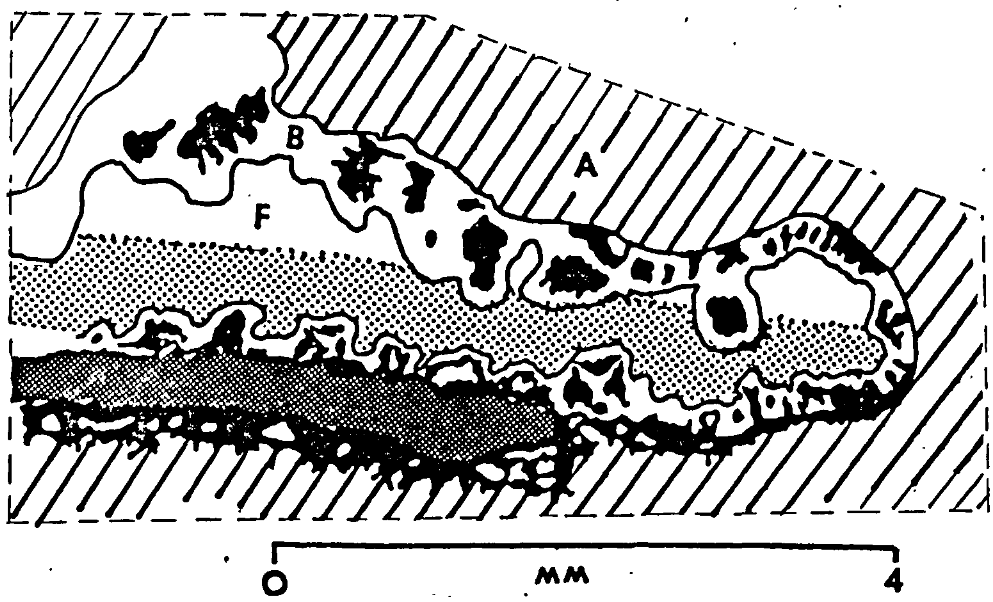
material occurs admixed with the upper 4cm of the bed (specimen 1472), and partly fills deep-penetrating tubular fenestrae as internal sediment, but does not line void walls and is itself overlain by a pure carbonate crystal silt (Plate 9, fig. E). Within the sediment the clay material is concentrated in vertical and horizontal sediment filled pipe (pedotubules s.s. of Brewer (1964, p. 236)) forms. Towards the upper bed surface the clay is homogenised. Notably fenestral cavities (both irregular and tubular) cross-cut pipes and localised zones of clay concentration indicating that clay admixing started before fenestra formation.

This admixing can be directly related to the development of a soil profile on the immediately overlying cycle top, and root-activity penetrating the partially lithified sediment (probably due to desiccation rather than cementation). This model indicates that some tubular fenestrae are produced by supratidal macrofloral colonisation.

Associated with the vadose cement fabrics described in sections 6.4.11 and 6.4.12. clay admixing from overlying 'wayboard' clays (associated with palaeokarsts) occurs. This takes the form of:
i) internal sediment replacement and infill of fenestral cavities, and
ii) partial replacement of the early cement fabric by cryptocrystalline clay. This latter example is illustrated by fig.50, and plate 8, fig. H . X.R.D. and S.E.M. with E.D.A.X. (- see Plate 10) analysis shows this clay to be dominantly?unorientated illite, with a subordinate mixed-layer illite-smectite tail to the 10Å peak. The replacement is preferential along growth zones within the fringing cements, and is dendritic in distribution, with thin clay fingers branching centripetally to the cavity but totally enclosed by the early cement. The degree of cement replacement decreases towards outer surfaces of the cement fringe. In the crystal silts, clay

Fig 50

Features of Fenestral Cavity with Microstalactites



A CALCISPHERE WACKESTONE (MA.1)
B Dripstone cements
C Illite replacements
D Clear Fe-calcite
E Geopetal silt
F Zoned to Fe-calcite. Spar druse

BASED ON MESOTHEM D5a D5b BOUNDARY AT FRONCYSYLLTE coats the silt grains, and is more pervasive towards the base of these silt-fills.

Associated with the illitic replacements are very clear ferroan calcite druses (zoned more ferroan to druse centres) (see Plate 9, fig. B). These areas are arranged towards margins of fenestrae, within or on the base of internal clay sediment fills associated with illitic infills, and intimately associated with clay replacements of cements.

Formation as a normal illuviation cutan (deposition from solution or suspension on natural surfaces within a soil) is improbable as this necessitates clay deposition synchronous with the carbonate cement. Clearly the replacement fabric precludes this. Brewer (1964) described quasicutans (subcutanic features that have a consistent relationship with natural surfaces within soils, but do not occur adjacent to or on these surfaces) that form by diffusion or precipitation internally within a soil plasma. Precipitation from solution is the most probable origin for the cryptocrystalline clay. Presumably the solutions moved along crystal interfaces by capillary transport, and preferentially precipitated the K-Al hydroxysilicate along certain zones (e.g. zones of high inclusion density, or fine micrite laminae, where higher crystal interface surface areas may have acted as a catalyst (?).)

The clarity and apparent void-fill cement nature of the ferroan calcite suggests that a degree of dissolution occurred prior to its precipitation. It is at this stage that silicate-bearing pore-waters may have partially replaced the carbonate cement. Ferroan calcites are often indicative of later stage evolved phreatic waters, although Richter and Fuchtbauer (1978) opined that argillaceous sediments may provide a ready source of iron to enter solution. As fenestral cavity infills are non-ferroan spar druse (zoned to slightly ferroan centres) they became cemented at an earlier stage than the dissolution-replacement ferroan calcite, which could only be precipitated from pore waters marginal to the now impermeable fenestral 'amygdale'.

6.4.14. Early diagenetic cherts

Lenticular, ~ 10cm thick chert bodies are rarely associated with MA.1. lithologies in the Tynant Formation (Chart A). As they are more pervasive in MA.2. lithologies, they are discussed in chapter 7.

6.5. PALAEONTOLOGICAL ASSOCIATIONS OF PERITIDAL SEDIMENTS.

6.5.1. Macrofauna

6.5.1.1.A. CNIDARIA. Within only MA.1.1 microfacies fragmented and abraded fragments of corals are rarely found in microscopic analysis. They are always partly micritised, and may contribute to the unidentified micritic peloids ubiquitous to all MA.1 microfacies.

B. PORIFERA. Chaetetid sclerosponges are a very rare component of Tynant Formation MA.1.1 microfacies. They are of bulbous form (~ 10cm diameter), are not preferentially orientated and often have eroded margins.

C. BRACHIOPODA. Large chonetid brachiopods (Daviesiella llangollensis)and rare Linoprotonia productids occur within MA.1.1. microfacies in the upper cycles of the Tynant Formation (e.g.s/m 197). They are usually in life position, with both valves intact, but do not occur in the abundance seen in MA.2 and MA.3. In the Sychtyn Member, athyrids of the Composita group occur locally in quantity in non-fenestrated MA.1.2a microfacies (e.g. s/m 1075). This Composita dominated assemblage is a characteristic feature of the Sychtyn Member.

D. MOLLUSCA. Gastropod dominated assemblages predominate in

MA.1 lithologies. Although their identification is mostly restricted to sectional characters, (due to extraction difficulties cf. Batten (1966)) approximate diversity counts show that up to 4 morphologically distinct forms occur in association. They are significant sediment contributors and often remain unfragmented. Bivalves are relatively scarce, and when present are usually small, recognised only petrographically.

Mollusc distribution, nevertheless is sporadic. Some units have relatively little gastropod material, whereas others contain greater than 5% volume (see appendix IV).

E. ECHINODERMATA. Occasional crinoid ossicles are present in MA.1 lithologies, but are normally abraded and micritised to varying degrees. Echinoid spines occur in MA.1.1., as a minor sediment contributor.

6.5.2. Macroflora

Carbonaceous films and poorly preserved plant and wood remains commonly occur scattered within and upon beds of MA.1 lithologies in both the Tynant Formation and Sychtyn Member (e.g. s/m 020).

6.5.3. Microfauna and Microflora

A. ALGAE.

Calcispheres. Throughout MA.1 lithologies calcispheres are significant sediment contributors (up to 26% by volume - Appendix II). They are diverse morphologically, but particular forms occur throughout the Asbian, especially *Pachysphaera* spp., *Palaeocancellus* spp., *Polyderma* spp. and *Quasipolyderma* sp . The thick walled *Vicinesphaera* and spinose *Parathurammina* are also common in MA.1 lithologies.

Red Algae. Red algae play a less important role in MA.1 than in other Asbian microfacies associations. cf *Epistacheoides* sp and *Ungdarella* sp are rare elements of MA.1.1 microfacies.

TABLE 5

STANDARDISED PETROGRAPHIC DIVERSITIES OF MICROFACIES ASSOCIATIONS.

SEDIMENTS	TOTAL DIVERSITY	MACROFAUNA DIVERSITY.	MICROFAUNA DIVERSITY	MICROFLORA DIVERSITY.	
Eglwyseg/Llynolys Formation M.A. 3	17. ± 4.3	5.9 ± 1.6	5.0 ± 2.5	6.1 ± 2.0	
Llanymynech Member M.A.3.	18.7 ± 2.6	6.0 ± 1.3	5.8 ± 3.4	7.0 ± 1.6	ASBIAN
Sychtyn M'br/Tynant Formation MA.3.	15.7 ± 3.2	3.7 ± 1.1	4.9 ± 2.1	7.1 ± 0.9	ASBIAN
Tynant Formation MA.2	15.0 ± 4.4	5.1 ± 1.4	4.1 ± 1.4	6.0 ± 2.6	ASBIAN
Sychtyn M'br/Tynant Formation MA.1.	7.7 ± 3.4	1.9 ± 1.9	2.0 ± 1.4	3.8 ± 1.3	ASBIAN
Trefor Formation MA.3	18.1 ± 3.0	6.3 ± 1.6	6.3 ± 0.8	5.5 ± 1.4	BRIGANTIAN
Trefor Formation MA.2	23.0 ± 2.5	9.1 ± 1.3	7.9 ± 1.2	6.0 ± 1.6	BRIGANTIAN

Sample population of 10 for each

Diversity counts are taken on 5cm^2 of thin section or acetate peel (approx. equivalent to the average thin section area). Each sample was analysed for 10 minutes to avoid bias of observation. Diversities represent diversity in morphological (microscopic) form of component clasts and do not represent true 'species' diversity.

Green Algae. Dasycladacean algae are represented by minor occurrences of Koninkopora sp.. They do not appear to be indigenous to this microfacies association. The beresellids, represented by Kamaenella denbighi Mamet and Roux, are ubiquitous to MA.1.1 and MA.1.4 but are rare in other MA.1 lithologies.

B. OSTRACODS.

Ostracods are ubiquitous to all MA.1 microfacies. They are only locally significant sediment contributors (in MA.1.1).

C. FORAMINIFERA.

Endothyrid foraminifera occur as a minor component of MA.1.1 and MA.1.4 microfacies. They may contribute to the micritic structureless peloids of MA.1.2. Archaediscoids and others have been recognised as trace components.

6.5.4. Diversities.

Standardised petrographic diversity counts (sensu Smosna & Warshauer 1978), including all organic components are shown in table 5. All groupings are low compared to the range of other microfacies associations (mean 7.7 ±3.4, maximum 14 population of 10). MA.1.1 microfacies have the highest average, bolstered by higher macro-faunal diversities compared to other MA.1. microfacies (maximum 6). Microfloral diversities have the highest mean (3.75 ±1.3), but are lower than MA.2 and MA.3 microfacies associations (see table 5).

6.6. PALAEOENVIRONMENT INTERPRETATION

Peritidal carbonate facies are well documented. Their recognition depends on the presence of:

1) Emergence indicators - fenestral fabrics; desiccation cracks and contemporary vadose diagenesis
2) A suite of sediments directly comparable to Recent peritidal environments
3) A faunal assemblage of 'restricted' character compatible with known peritidal environment limitations.

Generally '1' and '2' have less inherant ambiguity than '3'.

From the work of Black (1933) Cloud (1962) Shinn (1968); Shinn et al. (1969); Kendall and Skipwith (1969); Logan et al. (1974 b) and Hardie (1977) the anatomy of recent carbonate peritidal accumulations may be directly compared with the fossil record.

Recent carbonate tidal flats have three broad environment divisions: an outer zone that protects the tidal flat, either a beach ridge, a broad sublittoral platform, or a barrier island/reef/beach ridge complex; an intertidal flat, transected by tidal channels and gulleys, and some with tidal ponds; a supratidal zone dominated by algal marsh, or sabkha.

Three major controls affect Recent peritidal accumulations: the geometry of underlying bedrock, the climatology of the region, and the local hydrography.

Three major sedimentary processes operating on these deposits are the growth and sediment trapping by algae, storm sedimentation from suspension-charged waters and the growth and death of indigenous organisms, both animal and plant.

PALAEOENVIRONMENT SIGNIFICANCE OF MICROFACIES ASSOCIATION MA.1.

6.6.1. Calcisphere Mudstone to Packstone (MA.1.1)

The homogenous to poorly laminated nature of this lithology implies that mucillagenous algae (producing cryptalgal fabrics) were not significant sediment trappers, and/or that bioturbation was predominant. Vague mottling supports this latter statement (Moore and Scruton, 1957). Although fenestral fabrics are common in MA.1.1, their equally common absence supports a subtidal to intertidal formation. They commonly grade into the underlying MA.2 lithologies, implying a gradual lateral transition across their facies boundaries. The progradational model of sediment accretion demands that 'Walthers Law' (sediments deposited in laterally adjacent environments will be found vertically superposed in the geological record) is obeyed.

6.6.1.1. LIME MUD AND CALCISPHERES - SOME IMPLICATIONS Lime mud in most recent tidal flats is aragonitic. Its origin is still in dispute, but it is certain that calcareous algae and other skeletal components contribute substantially to its formation (Matthews, 1966; Stockman et al.1967, p.162). Neumann and Land (1975) estimated that in the Bahamas aragonite needles produced by green algae may account for more lime mud than is accumulating. The inorganic precipitation of aragonite mud (Bathurst, 1975, p.284-292) is likely to be a subordinate contributor compared with skeletal degradation.

The origin of Carboniferous lime mud promotes speculation. Much silt-grade bioclastic debris is resolvable with very thin sections, and with the S.E.M.. The ≤ 5μm micrite component may rarely be > 25%.

Bathurst (1975, p.278-279) emphasised that micritisation strengthens a skeletal grain, thereby producing carbonate sand preferentially. Penicillus, an extant green alga, is the most prolific lime-mud producer in both Floridan and some Shark Bay environments, but there is no direct evidence of a similar Carboniferous producer. However, calcispheres are dominant elements of MA.1 sediments. Their affinities have been disputed. Stanton (1963) suggested they were plant spores or reproductive bodies whilst Conil and Lys (1963), classified them within the foraminifera. Rupp (1967) likened them to reproductive cysts of extant acetabularian dasyclads, followed by Marszalek(1975) and supported by Wray (1977, p.104). Fragmented Asbian calcisphere cysts contribute to the degraded bioclastic silt and may be a significant lime-mud contributor, but the thallus of the hypothetical calcisphere plant may have provided (even if only partly calcified) a significant volume of lime mud. It is assumed that the calcisphere plants were subtidal epiflora, and therefore may have also played a major role in substrate stabilisation and mud trapping. Calcisphere distribution within the microfacies associations indicates that the plant most likely inhabitated MA.1. 1 substrates and that these sediments are therefore in part subtidal. Marszalek (1976) noted that Acetabularia 'calcispheres' do not survive transportation, although the distribution and apparent robustness of Carboniferous forms suggests they may have been transported considerable distances.

Calcisphere-dominated lithologies are known from similar palaeoenvironment settings, from the Lower Asbian of Ravenstonedale (this volume, p. 69), the Devonian of the Williston Basin (Wilson, 1967) Permian 'restricted' marine platforms (Wilson, 1975, p.222) and Cretaceous sediments (Lins, 1976 ; Flugel, 1978, p.226). Marzalek (1975) stated that Acetabularia calcispheres occur only in shallow protected waters with partly restricted circulation. Although the limitations of this palaeoenvironment model were opined by Lins (1976), there is a degree of correspondence between these Asbian calcispheres and their recent counterparts.

6.6.1.2. PALAEOECOLOGY AND TAPHONOMY OF MA.1.1. Extant cerithiid gastropods (e.g. Batillaria spp.) are dominant macrofauna elements of Recent carbonate tidal flats. They graze on surficial algal coatings, and are found especially within pelleted lime mud lithologies. They are particularly tolerant of fluctuating salinities and periodic emergence (Bathurst, 1975, p.198).

Multispecific gastropod assemblages characterise MA.1.1. It is the impoverished nature of other indigenous macrofauna which indicates that abnormal salinities were probable.

Periodic storms may have carried in other organisms, and fragmented indigenous shelly debris, bioturbation subsequently destroying any laminated deposit, but fragmentation of indigenous fauna was probably caused by both organism activity and oxidation of organic compounds (Matthews, 1966).

6.6.2. Unlaminated calcisphere-peloid grainstone/packstone (MA.1.2a)

The common association of fenestral fabrics with this microfacies (irregular and laminoid forms) indicates that it was mostly penecontemporaneously emergent. The peloid nature may be partly an early diagenetic ped formation, with associated growth of packing fenestrae. The presence of ?stacked, concave-up fenestrae sets indicates 'pustular' algal mats probably formed and were grazed by gastropods. Many of the peloids resemble larger calcispheres in size and shape. Logan, Hoffman and Gebelein (1974a, p.152) noted that pellets are common grains in cryptalgal sediments, formed by the diagenetic alteration of algal-trapped carbonate grains to cryptocrystalline micrite, by endolithic algal activity (Bathurst, 1966). A similar process is envisaged for formation of MA.1.2a, as an intertidal deposit. Bioturbated pelleted lime muds occur subtidally on Andros (Ginsburg and Hardie, 1975, p.206) and provide a recent analogue for non-fenestrated MA.1.2a.

6.6.3. Laminated calcisphere-peloid grainstone-packstone (MA.1.2b)

Similar deposits occur in Bahamian peritidal environments prone to intermittent flooding and exposure (Ginsburg and Hardie, 1975, p.206). These sediments occur on levee crests and backslopes, inland algal marshes, and beach ridges (Shinn *et al.*, 1969). The presence of penecontemporary dolomites associated with this microfacies is significant. Hardie and Ginsburg (1977, p.59) described similar graded peloid micrite couplets on levees. The lack of evidence for presence of penecontemporary gypsum implies that dolomitising pore-waters were < 65‰ saline and therefore dolomitisation could occur at relatively low Mg/Ca ratios(< 1 0 according to Folk and Land, 1975). The presence of well-formed early cement rhombs as well as the micrite grade carbonate replacement remains an enigma, although Kocurko (1979, p.211) noted the formation of c 60μm rhombic dolomite cements by spray zone brine seepage and Gebelein (1978) advocated pervasive dolomitisation in meteoric-marine tidal flat mixing zones.

Graded laminae form by deposition from suspension during flooding, but its preservation relies upon non-bioturbation. Schenk (1967) described laminated non-dolomitic pelleted, and dolomitic bituminous sediments formed by storm-tide activity from Mississippian strata, associated with intraclastic laminae. MA.1.2b formed by periodic flooding of the tidal flats during high tides or storms, (Ball *et al*, 1967) with erosion of lithified carbonate crusts and contemporary resedimentation as intraclasts. Sedimentation may in part have been induced by algal films binding the particles (their presence inferred from laminoid fenestrae and associated micritic cryptalgal laminites). Hardie and Ginsburg (1977, p.94) found that no astronomic tides carried sufficient suspended sediment on to the tidal flats to affect sediment deposition. Colonisation by tubular fenestrating organisms, in this instance possibly supratidal flora, occurred subsequently.

6.6.4. Cryptalgal laminites (MA.1.3.)

Fine micritic laminae, with much ≤ 20μm detritus, laminoid fenestrae, and associated disruption with intraclast breccias indicate formation by binding activities of mucilagenous algal mats in a high intertidal to supratidal environment (Logan 1961, Black 1933).

The presence of fine detrital quartz rich laminae suggest that wind may have been a minor transportation agent (?on exposed high tidal flats.)

Recent domal smooth mat equivalents to the LLH stromatolitic forms of this sequence occur in lower intertidal zones of Shark Bay (Logan et al 1974a).

6.6.5. Laminated beresellid calcisphere mudstone-packstone(MA.1.4)

This microfacies is characteristically non-fenestrated. It lacks a stenohaline faunal assemblage. It is here interpreted as deposited in a subtidal environment of restricted circulation, (sensu Wagner and Togt, 1973) and without any infauna to disrupt the lamination. Lamination may reflect either periodic beresellid blooms or tidal/storm sedimentation.

6.6.6. Clast supported conglomerates.

Associated with fenestrated MA.1 microfacies and S.D. surfaces (chapter 11), these sediments resemble breccia intraclast pavements, but differ in not having a dominance of 'platy' clasts (Hagan and Logan, 1974, p.101). A pre-requisite for their formation is a lithified, or partly lithified and coherent source sediment. Although many recent intraclast breccias form by in-situ fragmentation of the substrate and little transport,(due to cementation expansion, growth of gypsum, and desiccation), a degree of transport is likely for these Asbian conglomerates due to their subspherical rounded gravel and cobble form. Hardie and Garrett(1977) described conglomerates of exhumed cohesive but unlithified and homogeneous pond sediments (pellet muds) formed at the base of beach cliffs (c.40cm relief) at the upper level of intertidal beaches on Andros. Hardie and Ginsburg (1977, p.88) attribute the erosion of this beach cliff to 'wave action during storms', but note that the process is enhanced by the 'Swiss Cheese' texture of the sediment produced by burrowing crabs. As explained in section 11. 1 , S.D. surfaces, that underlie such conglomerates, have varied burrow features that may further help the formation of these analogous sediments.

Burgess (1978) described storm deposits from Fiji, 20 to 50cm thick, fining upwards, including 'bioclasts' up to 40cm diameter in a matrix of 'highly variable allochems.' No fining upward structure is apparent in these Asbian deposits.

6.7. EVIDENCE OF PALAEOTIDAL RANGE

Klein (1971) provided a model for identifying palaeotidal range in clastic tidalite sequences based upon recognition of low tidal flat and high tidal flat sediments in 'fining upwards' progradational sequences. According to this model, palaeotidal range was estimated as the stratigraphic thickness between these tidal subenvironments. The exposure index model of Ginsburg and Hardie (1975) provides a basis for similar carbonate palaeotidal range analysis.

The presence of prograding fining upward peritidal cycles within the Lower Asbian, allows application of Klein's concept as a palaeotide range guide line. This, however, neither takes into account subsidence nor eustatic influences over the period of progradation, but includes the joint influences of both meteorologic and astronomic tides.

Palaeotide Prediction:

ASTRONOMIC PALAEOTIDES. Prediction of astronomic palaeotidal range from palaeogeographic and chronological constraints may, at present, only provide gross guidelines. According to de Klein (1971) and Schopf (1980) the spectrum of palaeotidal range has not significantly altered through geological time, even though the effects of the Moon may have changed considerably (Rosenberg and Runcorn, 1975).

Small enclosed epicontinental seas have (at present) low tidal ranges (Schopf 1980) of the order of centimetres at maximum.

The palaeogeographic constraints of the Craven Basin to the north and east of the study area, suggest that the astronomic palaeotide contribution was the order of centimetres rather than metres.

Meteorological tides were, therefore, probably of greater significance on sedimentation than astronomic tides.

METEOROLOGICAL PALAEOTIDES. The major extent of the 'Craven Basin' (fig. 3) is about 100km extending north east from the study area.

The two peninsulae of St. George's Land, (Cyrn y Brain and Berwyn 'axes') would have provided minimal protection from north easterly storm winds, but would themselves have borne the most severe storm wave effects from winds in these quarters.

Storm surges, caused by 'wind stress' would produce meteorological 'tides' in coastal areas. Welander (1961) produced an equation defining this storm surge amplitude which was applied by Hardie and Ginsburg (1977) to the tidal flats of Andros (Bahamas).

Whilst the overall fetch of the Craven Basin was approximately 100km, the shallow shelf region along the flanks of St. George's Land was probably less than 15km wide, at least during the early Asbian. These constraints suggest that at wind speeds of 20m / sec. (a tropical storm by Recent Andros standards) meteorological tides would be produced of less than 1m amplitude. (cf. Hardie and Ginsburg (1977)).

Evidence of palaeotidal range.

Sediment characters indicative of a high palaeotidal range are absent within MA.1 sediments. The lack of tidal channels / creeks within the peritidal sediments suggests that microtidal ranges dominated (< 2m tidal range) (Walker 1979, p. 58.).

Estimates of palaeotidal range using de Klein's method are open to problems of recognition both of multiple cyclicity upon emergence (chapter 12) and of specific intertidal zone fabrics. The fenestral palaeoenvironments discussed in section 6.5., suggests that there is a degree of preferred sequential distribution of fenestral fabrics, although palaeoenvironments of fenestrae are poorly defined.

Figs 9a & 43 indicate that in stratigraphic profile, fenestral fabrics are variably associated from thin dense fenestral levels to thick metre-scale sequences containing scattered fenestrae.

The occurrence of dense medium to small, irregular fenestrated units is, however, ubiquitous throughout the Tynant and Whitehaven Formations. These most distinctive levels vary from 5 to 40cm in thickness. They indicate an absolute minimum for palaeotidal range, whilst absolute maximum ranges must be based on total thickness of fenestral fabric sequences that are almost exclusively < 2m thick. Apart from upper minor cycles of Mesothem D5a, all MA.1 units themselves are ≤ 2m thick, but many of these lack fenestral fabrics and were probably in part of shallow subtidal origin.

High energy erosive and deposition events within MA.1 units are relatively uncommon, (Fig. 43) suggesting large scale storms were rare tidal flat modifiers, although the fine laminated MA.1.2 sediments may be analogues to Recent Andros storm deposited sediments on a frequent (seasonal) scale.

6.8. SUMMARY AND CONCLUDING REMARKS.

The microfacies within Tynant Formation and Sychtyn Member deposits are very comparable, and are stratigraphically arranged in a similar manner, in similarly structured minor cycles (see section 2. 1. 1 and fig. 10). The sediments are readily grouped into four microfacies .

The Calcisphere Wackestone microfacies association (MA.1) is the characteristic upper phase of these cycles, is dominated by calcisphere mudstone/packstone, (MA.1.1) and includes the 'Porcellanous limestones' and 'calcite mudstones' of other workers. Other minor microfacies recognised are calcisphere-peloid grainstone / packstone (MA.1.2): cryptalgal laminite (MA.1.3): and laminated beresellid-calcisphere mudstone-packstone (MA.1.4). These sediments represent a progradational'tidal flat' phase of deposition. Penecontemporary dolomites are associated with these microfacies only and bladed, fibrous and micritic isopachous early cements line some fenestral cavities, especially near erosion surfaces and cycle boundaries.

Laminoid, irregular and tubular fenestral fabrics are recognised associated with MA.1, along with 'packing fenestrae' defined here as a fenestral fabric that imparts or enhances a grainstone texture to the sediment. Markovian analysis of fenestral stratigraphic distributions show that at $>90\%$ confidence there is a cyclicity in fabrics. The fenestral fabric relationship sequence is: tubular; large irregular; small irregular; laminoid. In exceptional cases, associated with omission surfaces, refenestrated fenestrae occur. Tubular fenestrae were formed mainly by burrowing infauna, although evidence of ?plant rhizome systems indicates that they may in part be supratidal. Large irregular fenestrae represent low intertidal to shallow subtidal algal mat diagenesis, whilst medium and small irregular fenestral fabrics apparently formed intertidally. Laminoid fenestrae suggest high intertidal (with rare low LLH stromatolites and supratidal environments. Packing fenestrae are associated with irregular ped structures, and their formation mechanism may be similar to the 'planes' common in soil plasmas. No evidence of desiccation cracks have been found. These fabric associations attest well to the shoaling cycle model.

Along with the early cements, are grain supported, lime-mud supported and crystal silt internal sediments indicating a complex history of early lithification and vadose erosion. At two localities on the Mesothem D5a/b boundary microspeleothemic and curtain cements indicate early meteoric vadose cementation. Clays both replace and are transported into MA.1 sediments and cements, where they underlie prominent palaeokarst surfaces, and are indicative of soil forming processes. Penecontemporary dolomite cements and replacements are depicted as having formed in mixing zone environments on tidal and supratidal flats, without any evidence of an evaporite association, as in Recent Andros Island 'dolomites'.

(In recent sediments, gypsum precipitates above 65‰, salinities although high alkalinities (induced by organic decomposition) may retard this.)

MA.1 sediments have a low petrographic diversity (~8 ±3), and apart from molluscan material, macrofaunal clasts and indigenous benthos are rare. Non-molluscan elements mostly show a degree of micritisation and are highly abraded , indicating transportation, relative to the few indigenous elements.

Calcisphere wackestones (MA.1.1) are interpreted as subtidal to intertidal in origin (depending on presence or absence of fenestral fabrics). The calcisphere plant may have both acted as a shallow subtidal sediment baffle and stabiliser, and even if its thallus was only partly calcified, could have contributed significant volumes of lime mud. The calcispheres (reproductive algal cysts) themselves contribute substantially to the sediment (up to 26% by volume).

Calcisphere-peloid grainstone/packstones (MA.1.2) are interpreted as intertidal to supratidal deposits, with algal mat fabrics, intense mm lamination locally (MA.1.2b), and penecontemporary dolomites, that may be direct analogues with Recent supratidal storm laminites. Centimetre-sized intraclasts are commonly associated.

Cryptalgal laminite (MA.1.3) is a minor microfacies, of fine micritic or detrital-clastic rich laminae, and is normally interbedded with other MA.1 microfacies, and may represent a high intertidal to supratidal environment.

Laminated beresellid-calcisphere mudstone-packstones (MA.1.4) are rare subtidally deposited sediments, mostly occurring in the Sychtyn Member, and indicative of a faunally restricted environment.

MA.4.1 are normally associated with MA.1 units, and may represent erosion of coherent or partly lithified sediments, either by tidal action, or more likely, storms. Similar sediments occur in Recent Andros beach environments.

Low centimetre scale astronomic tidal ranges are suggested.

7. SUBTIDAL DEPOSITS OF THE TYNANT FORMATION AND SYCHTYN MEMBER

Microfacies represented within the subtidal sediment suite of the Tynant Formation and Sychtyn Member;

Argillaceous Alga Packstone Microfacies Association (MA.2)
MA.2.1. Beresellid wackestone to packstone
MA.2.3. Calcareous shale

Alga Peloid Grainstone Microfacies Association (MA.3)
MA.3.1. Alga Peloid grainstone
MA.3.2. Bioclast grainstone
MA.3.3. Peloid grainstone

Bioclast Peloid Rudstone Microfacies Association (MA.4)
MA.4.1. Bioclast Peloid Rudstone
(Matrix Supported Intraformational Conglomerate (L.7.2))

7.1. DISTRIBUTION AND FIELD CHARACTER OF MA.2.

Alga packstones are dominant lithotypes occurring towards the base of minor cycles with MA.1. upper phases. They also occur within the Llanymynech Member, the Eglwyseg and Llynclys Formations and are dominant Trefor Formation lithologies. In the Tynant Formation they may contribute up to 50% of the measured section (see chart A), but are minor microfacies of the Sychtyn Member.

Characteristically they are thin (5 to 80cm) parallel to wavy bedded lithologies, often dark grey to black with variable argillaceous content. They grade into calcareous shales, which are here included within MA.2., into calcisphere wackestones (MA.1.1) and into MA.3.1. Typically they may be interbedded with or underlie MA.3 lithologies. They are always associated with a stenohaline faunal assemblage.

7.2. PETROGRAPHY OF MA.2 SEDIMENTS

7.2.1. Petrography of MA.2.1. Beresellid Wackestone to Packstone

This is the dominant microfacies within the lower cycles of the Tynant Formation, but occurs to a lesser extent in the Sychtyn Member. With increasing proportion of calcispheres and relative decrease in other algae MA.2.1 grades into MA.1.1.

The sediments are variably argillaceous and grade into MA.2.3. They may be millimetre laminated and break into 'flags' along partings, or more commonly homogeneous. Clast sizes are variable, but normally $\leq$ 1mm. Rarely they are ripple cross laminated (Plate 11, fig. A).

Compositionally, macrofaunal elements are minor but variable components. Brachiopods, crinoids, tabulate corals and gastropods may be each a locally significant sediment contributor (up to 10% by volume). Echinoid spines are ubiquitous.

Algae are the most significant sediment contributor (av.76 $\pm$ 12% of recognisable biogenic components), and primarily the beresellid Kamaenella denbighi in the Tynant Formation, that can locally account for 65% volume of rock. Girvanella thalli, dense along certain laminae probably were sediment binding agents (e.g. s/m 006, Tynant Formation, Tynant). Both green and red algae are diverse (see section 7. 2.6.3.).

At times foraminifera are major components. Ostracods are ubiquitous. Crinoid fragments may reach 4mm and be relatively unabraded.

Micritisation of clasts is not common. Most fragmentation is high and rounding is low. Non micritised microborings (c. 20µm diameter) extend into a few bioclasts, (Plate 11, fig. D). (e.g. s/m's 243, 249, Tynant Form'n).

Macroflora carbonaceous films and rare pinnules occur along bedding planes.

Bioturbation and burrow-fill structures are rare, but visible as areas of grainstone texture or of greater mud content

PLATE 11 <u>Argillaceous Alga Packstone Microfacies Association</u>

A Ripple laminated MA.2.1. s/m 004. Tynant Formation, Tynant, G. R. SJ 22034535.

B Beresellid packstone MA.2.1.

. Small tubular elements within the matrix are <u>Kamaenella denbighi</u>. s/m 248. Tynant Formation.

C Calcareous Shale (MA.2.3) s/m 016. G.R. SJ 22064534

D Clast of pseudopunctate brachiopod pervaded by branching borings s/m 243, Tynant Formation G. R. SJ 21964572.

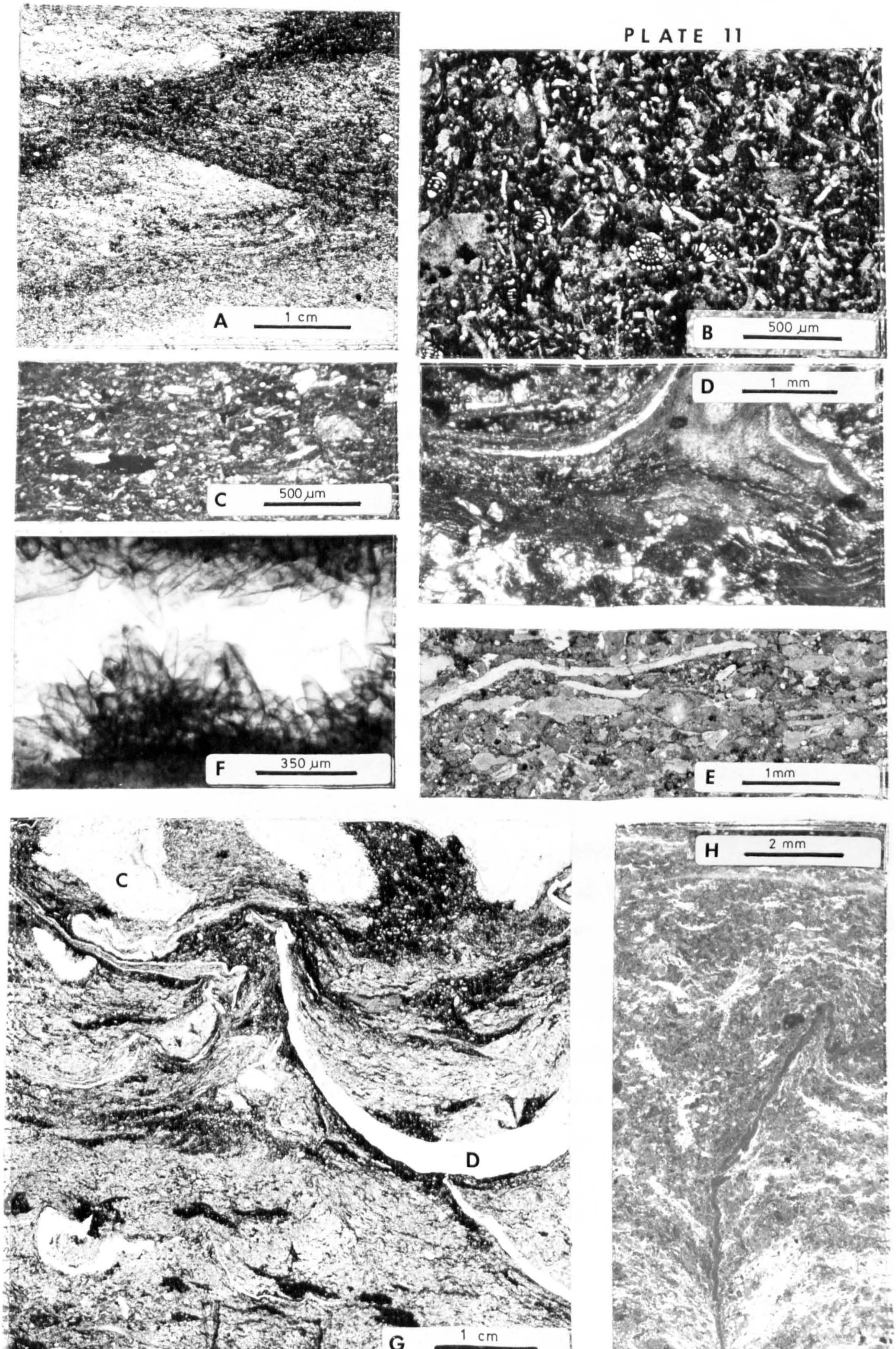
E Microstylolitic 'compacted' bioclast packstone, MA.2.2. The original texture may have been grainstone. s/m 195. Tynant Formation G. R. SJ 23424947.

F Acicular cement fringe to lumen cavity of chaetetid sponge. s/m 2525, Tynant Formation, G. R. SJ 21974581.

G Beresellid packstone with chaetetid (C) and <u>Daviesiella llangollensis</u> (D). Note presence of thin valved <u>Linoprotonia</u> sp within the sediment and burrow forms (dark) rich in argillaceous material. s/m 1041., Tynant Formation G.R. SJ21954572

H S/m 1041. Compaction feature around vertical brachiopod clast . Note turning of laminae to subparallel the clast near its margins. Total compaction ~50%.

PLATE 11



within the packstone (e.g. s/m 1041, Plate 11, fig. G). Homogenisation reflects pervasive bioturbation, and rare Zoophycos imparts a secondary distinctive lamination to the sediments. Whole fossil packstone centimetre layers occur intermittently, containing indigenous biota with partly fragmented brachiopods (mainly Linoprotonia and athyrids) and partly fragmented corals, echinoderms and molluscs in a typical MA.2.1. matrix.

Microstylolitisation is most pervasive in more highly argillaceous sediments., and is often accompanied by extensive micro/pseudosparitisation of lime mud matrix. Compaction causing clast fragmentation and distortion (as exemplified by ellipsoid calcispheres) occurs in most argillaceous specimens. These argillaceous sediments show little matrix except silt-grade bioclastic debris, probably derived from fragmentation by burial compaction. S/m 1041 shows clearly the effects of fragmentation with compacted Chondrites burrows and vertical brachiopod fragments surrounded by compaction 'envelopes' (stylolite defined) indicating about 50% reduction from original thickness (Plate 11, fig. H). Such compaction enhances the apparent total bioclastic contribution as the mud matrix is apparently more pressure compacted. This explains the high clast percentage in many MA.2 samples point counted (Appendix II). Significantly, some apparent MA.2.1. sediments contain both a minor proportion of recognisable cement and even primary uncemented internal fossil voids (e.g. s/m 026). Sychtyn member MA.2.1 microfacies are rarely appreciably argillaceous, and as such have less compaction phenomena.

Cements are non-ferroan calcite, and echinoderm fragments often have syntaxial rims.

With decreasing compaction, increasing cement and reduction in mud grade matrix MA.2.1 grades into MA.3.

7.2.2. Petrography of MA.2.3. Calcareous shale.

This is a common lithology within the Tynant Formation, but is rare in the Sychtyn Member. Purer shales or mudstones (L.6) are greys greens ochres and maroons. MA.2.3 is compositionally similar to MA.2.1, but has reduced quantities of recognisable algal debris (e.g.016 Appendix II. Their bioclast composition plots in a similar composition field to MA.2.1 (see ternary diagram, fig. 18). They have brachiopod (Daviesiella llangollensis and Linoprotonia cf hemisphaerica) dominated macrofaunal assemblages commonly associated. Rarely (e.g. s/m 241) foraminifera (mostly Endothyranopsis) contribute up to 20% of the sediment

Compaction features (partly microstylolitic) are very prominent, enhanced by compaction of the argillaceous component. X.R.D. shows that illite is the dominant clay, whilst mixed layer illite-smectites and kaolinite are weak (comparatively feeble 001 peak) or absent (Appendix III).

7.2.3. Bed Morphologies of MA.2

5 to 80cm beds of MA.2.1 are defined by flat or smoothly undulose bedding planes. These beds are often separated by centimetre shale partings or MA.2.3, with grading contacts between, or have 'multi-storey' stacking. Thin beds may be traced for hundreds of metres across outcrops, but in both the Tynant Formation and Sychtyn Member, their use as correlation markers is hindered by their laterally variable bed thickness and their commonness in the succession (particularly the Tynant Formation).

7.2.4. Lamination in MA.2.

As previously mentioned, in MA.2.1 microfacies, parallel millimetre scale lamination is common, and ripple lamination is rarely visible.

The millimetre bedding-parallel laminae are defined by both presence and absence of grainstone texture, and variation in bioclastic components. In s/m 243 (Tynant Formation, G.R. SJ 21954570) centimetre lamination occurs within MA.2.1 from fine algal packstones to brachiopod coral floatstones, defining horizons rich in whole and fragmented macrofauna.

Low angle cross lamination (unidirectional) occurs at a number of localities in the Tynant Formation, either as a primary bedform, or (see fig.37) as a sediment drape over pre-existing topography (presumed palaeokarstic). At G.R. SJ 22114636 (Tynant Formation) gently sloping foresets (c. 10° dip) define a bedform relief of 60cm. This lamination is defined by centimetre grainstone-packstone couplets.

Ripple cross-laminae (e.g. s/m 004 G.R. SJ 22024535) of amplitude c. 2cm and wavelength c. 10cm. have ?opposed foresets, and scoured set bases suggesting formation by wave processes (de Raaf et. al.,1977).

7.2.5. Early Diagenetic Features Associated with MA.2.

7.2.5.1. TABULAR AND NODULAR CHERTS. Cherts occur at a number of horizons within the Tynant Formation, mostly within MA.2.1., but occasionally within MA.1 and MA.3 lithologies. They vary from laterally persistent tabular forms (Plate 7, fig. D) (c. 10cm thick) to small locallised elliptical nodules. The latter commonly have a marked colour fringe around them and rarely envelop fossils. The nodules are dominantly equant microcrystalline quartz (< 20um) with irregular lenticles of equant undulose megaquartz (terminology after Milliken, 1979). Length fast chalcedonic silica occurs in bioclast cavities (e.g. Chaetetid lumina). There is no zonation of quartz fabric within the nodules. The outer pale fringe comprises two distinct zones: i) An inner 1 - 4 mm neomorphic pseudospar calcitic layer without recognisable clasts; ii) an outer centimetre zone of partly silicified clasts within microdolospar and microspar (5-30μm) matrix, leaving residual calcitic (dominantly brachiopod) clasts and microspar matrix. External to this dolomitic fringe, along the chert horizon, partial silicification of bioclasts continues (s/m 2451-iii) but all clast types are equally affected.

Within the chert nodules is an outer zone inward of the fringe rich in irregular calcitic 'clasts' and euhedral dolomite rhombs that are zoned to ferroan margins, many with microquartz-replaced centres. These decrease centrally within the nodules. In comparison the tabular cherts are milky with 10μm dolomites and calcitic clasts scattered throughout. Figure 51 outlines this petrographic zonation.

Evidence of early formation of the chert nodules is indicated by:

1) a small lenticular chert nodule orientated subvertically with ?lamination disruption around it (Fig. 52 .) (s/m 2523, G.R. 21944560), suggesting disruption (by bioturbation at a penecontemporary stage?).

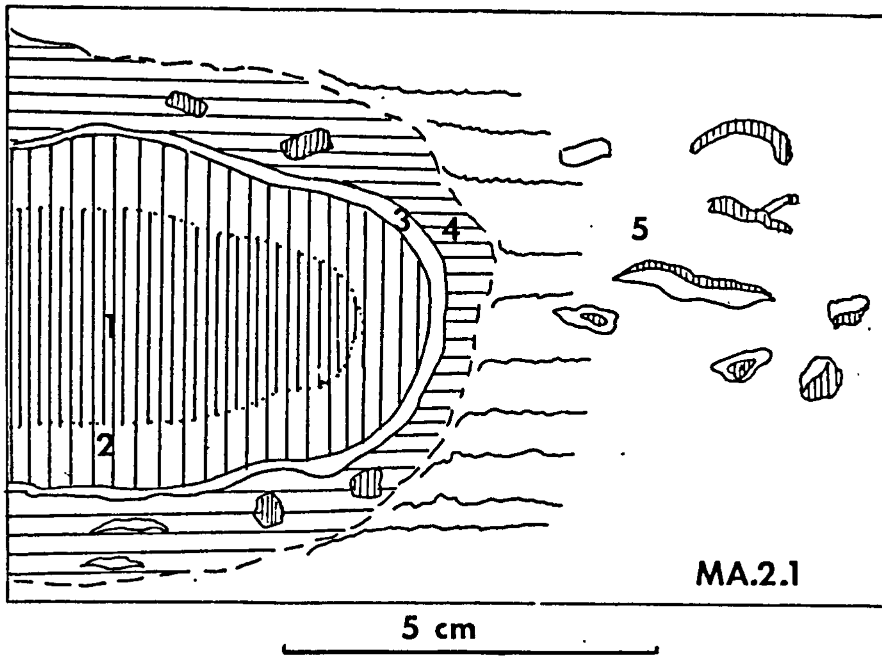
2) Growth of 'botryoidal' chalcedonic silica into primary (chaetetid lumina) bioclast cavities, indicating growth before final cementation.

3) Formation before much of the stylotitisation and compaction is shown by lamination and stylolites expanding towards the nodules (see fig. 51).

4) Presence of relict siliceous spicules within chaetetid sclerosponges only where associated with chert nodules indicates

CHERT NODULE PETROGRAPHY– MA.2.1

Fig. 51



KEY

1. Microcrystalline quartz and equant undulose megaquartz. Rare, partly silicified euhedral dolomite.

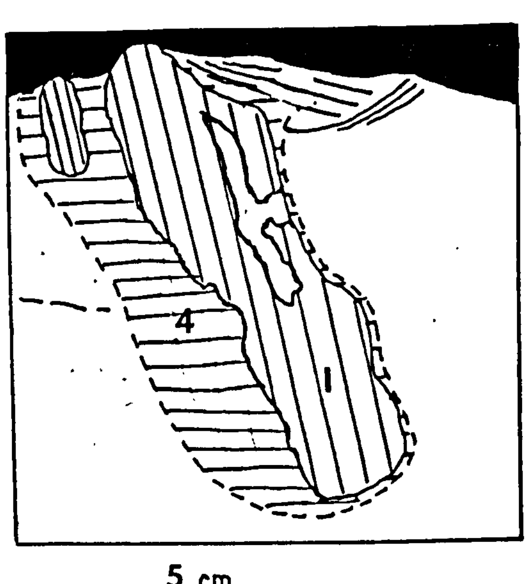
2. Microcrystalline quartz and equant undulose megaquartz. Euhedral dolomites prolific ($\leq 300\mu m$), many with centres replaced by microcrystalline quartz, and with ferroan margins.

3. Pseudospar (neomorphic fabric) (20–60µm) low Fe calcite.

4. Dolomite and calcite microspar, with relict calcitic bioclasts. Echinoderm clasts preferentially silicified.

5. MA.2.1 Beresellid packstone . Many clasts partly silicified. Microstylolites circumscribe the chert nodule.

FIG 52



SUBVERTICAL CHERT NODULE–TYNANT FORMATION

This chert nodule is atypical in its orientation – others are sub-horizontal. Note dolomitic fringe (4) encompasses ?partly disrupted surrounding sediment – indicating some growth post ?reorientation.

that silicification occurs prior to and during active dissolution of the primary opaline silica of the spicules, thereby retarding the dissolution process. As opaline silica readily dissolves in many carbonate shelf micro-environments (e.g. Hartman and Goreau 1970, Land 1976) this suggests silicification was early (Gray 1980).

Implications The presence of euhedral dolomite associated with and replaced by early diagenetic cherts promotes speculation. This has been recognised as a common association (Hey 1956; Wells, 1956; p.185; Meyers 1977). It implies that the dolomitisation is equally early, and that the two processes (silicification and dolomitisation) are intimately associated. Once silicification has enveloped the dolomite rhombs, their geochemical system is virtually closed.

Biogenic opaline silica was present as sclerosponge spicules that were actively dissolving in the sediment. However, the quantities involved are minimal compared to the degree of chertification. Only rarely do chert nodules form around chaetetid colonies. The sediments are argillaceous, and as pointed out by Meyers (1977) leaching of argillaceous sediments appears a favourable silica source. From isotopic evidence, Knauth (1979) proposed a model for chert formation in the marine-phreatic and vadose mixing zone. He stresses the intimate association of dolomites with cherts as evidence of their mutual mixing zone formation. Here a low pH increases calcite solubility but lowers that of silica (Correns 1950) as CO_2 partial pressures decrease in a closed system (Wigley & Plummer, 1976). The associated dolomites may represent early Dorag type dolomitisation (Badiozamani., 1973).

7.2.5.2. EARLY ACICULAR CEMENTS. Within bioclast primary cavities in MA.2.1 an irregular acicular fringe of clear calcite is rarely developed (Plate 11,fig. F). This cement texture has only been noted in partly silicified chaetetid colonies. The needles are ⩾ 100μm by 20μm, growing centripetally from bioclast walls and / or grading into equant 20μm calcite silts. Centripetally they have sharp terminations against coarse, clear void-filling spar.

Pingitore (1971) described acicular (early) cements from primary pore space in Pleistocene to Recent corals, with similar morphologies. Pingitore concluded that this cement was aragonitic, and precipitated in the marine environment (pre-uplift). He noted that cements associated with recrystallised (calcitised) coral skeletons left no trace of precursor acicular aragonites. Taylor and Illing (1971) described similar cement (aragonitic) textures recrystallising to magnesian calcite silt.

The association with chaetetid skeletons of probable original aragonite mineralogy (Gray, 1980) is significant.

7.2.6. Fossil Associations with MA.2 Sediments

7.2.6.1. A. CNIDARIA. In the Tynant Formation tabulate corals *Syringopora* and *Cladochonus* are common, with rarer fasciculate lithostrotiontids (*Lithostrotion* sp.nov.) subordinate. They are invariably associated with brachiopod dominated assemblages. The colonies are normally entire, always have a degree of crushing, and often orientated with their growth axes parallel to bedding planes, indicating slight ?post-mortem reorientation. Coral bioclasts are common only in the scarce coralliferous horizons. These horizons are of limited lateral extent, and occur both in MA.2.1 and MA.2.3 sediments. Colonies are scattered, seldom < 150cm apart. They are never profuse enough to merit the term 'biostrome'.

Similarly in the Sychtyn Member Syringoporoid-fasciculate lithostrotiontid (Lithostrotion sp. nov.) associations occur, but are infrequent. Syringoporoid-chaetetid associations, on the other hand, are prolific and discussed in section 7.3.6.1 .

B. PORIFERA. In the Tynant Formation MA.2 microfacies, chaetetids are present in minor quantity (slightly more abundant than the coral macrofauna) and occur as whole or fragmentary colonies up to 12cm diameter in various states of compaction fracture.(Gray, 1980).

In the Sychtyn Member chaetetids (C.(Boswellia) primarily) are common in both MA.2 and MA.3 microfacies, and are always associated with syringoporoids. They form laterally discontinuous biostromes of about 1 colony/m in horizontal line-section. Their growth morphologies are bulbous but rarely pyriform. They occasionally encrust brachiopod valves, and are mostly in growth-orientation.

C. BRACHIOPODA. The most common macrofaunal elements in both Tynant and Sychtyn MA.2 microfacies, brachiopods are nevertheless not diverse in comparison to MA.2 microfacies of the Trefor formation.

In the Tynant Formation Daviesiella llangollensis (a gigantid chonetid) is the most readily recognised brachiopod, occurring in profusion along horizons (e.g. fig. 53), but mostly of variable orientation. Rarely are greater than 70% of their valves in life orientation. Most commonly the daviesiellids occur within thin MA.2.3 horizons on minor cycle bases (see chart A). It is in these calcareous shales where the greatest proportion of specimens in life orientation, with both valves intact, are found. Daviesiellid "sectional distribution densities" (S.D.D." new term, defined in Fig.53) do not go below unity in local dense accumulations, and average out above 3 over outcrop (fig 53). Sediment compaction reduces this from the initial burial value. Linoprotonia cf hemisphaerica is the most common productid, occurring in daviesiellid dominated assemblages or discontinuous along horizons as the dominant faunal element. Most are found in life orientation (e.g. s/m 1041), with ventral and dorsal valves attached. Others show little sign of significant transportation. Their ventro-lateral and trail spines partly stabilised their orientation. When in dense association, their valves are within millimetres of lateral contact with each other. Spiriferids and Athyrids are minor but ubiquitous brachiopod components (see appendix IV).

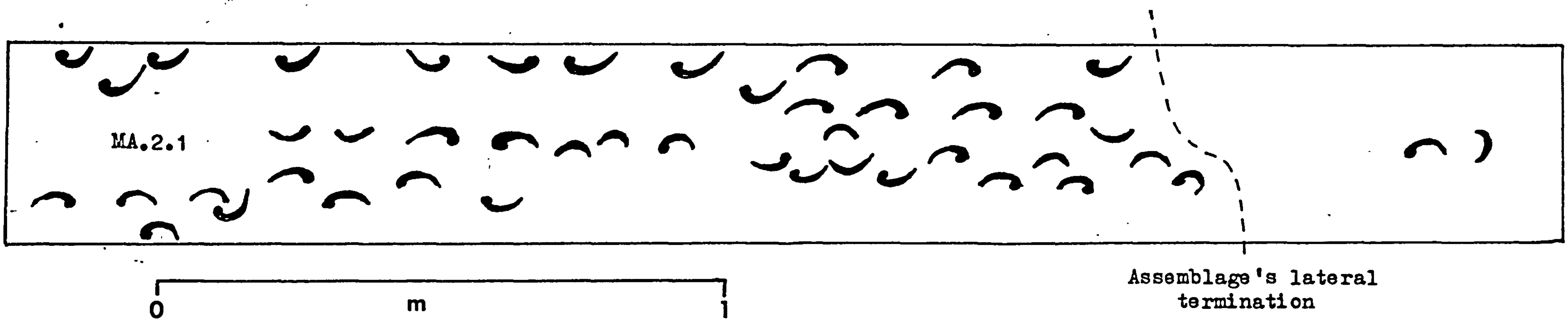
In the Sychtyn Member Daviesiella Llangollensis occurs only within the lowermost strata. Linoprotonia is a dominant element , but Composita sp. is the most abundant, in both MA.2 and MA.3 microfacies. In MA.2 they are mostly fragmentary, disarticulated, and are variably compaction-fractured (e.g. s/m 2150). Their prismatic microstructure easily disaggregates, making locally a significant sediment contribution (Section 9.3.1.2.).

D. ECHINODERMATA. Rare non-micritised, little abraded crinoidal packstone laminae indicate that local crinoid 'stands' may have been present. Although echinoid spines are common in MA.2 microfacies they never reach the proliferation shown in Llanymynech Member MA.2 microfacies (see Section 8.1.2.1).

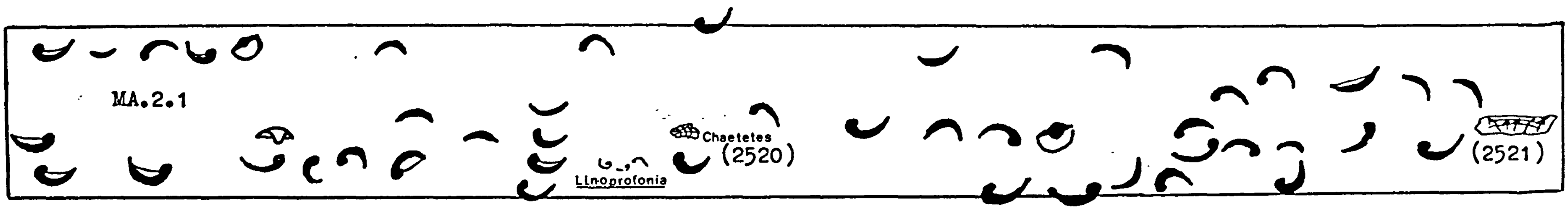
7.2.6.2. MACROFLORA Plant debris occurs commonly within fossiliferous L.6.2 &MA.2.3 microfacies of the Sychtyn Member, and rarely within the Tynant Formation, indicating a proximal subaerial source. This debris is normally fragmented fronds and pinnules, but preservation

Lateral log of a *Daviesiella* bed in the Tynant Formation

A Tynant Ravine (SJ 21964592)



B Eglwyseg Escarpment (SJ 22152638)



SECTIONAL DISTRIBUTION DENSITIES: (S.D.D.)

Defined as the ratio of 'boxed' sectional area of a brachiopod to its average share of quadrat area i.e.

$$S.D.D = \frac{O}{A.N}$$

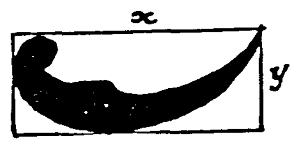
Where O = total quadrat area
A = average 'boxed' sectional area
N = number of brachiopods in quadrat area (ventral valves only).

S.D.D for bed 'A' = 5.1

S.D.D for bed 'B' = 4.0

% ventral valves in life orientation for bed 'A' = 44%

——— " ——— 'B' = 48%



'Boxed' sectional area (A) = x x y

Fig 53

10.1.2. Petrography of MA.2.1. Alga packstone to wackestone

This microfacies resembles MA.2.1 of Asbian sediments, except that beresellid and tubular/septate algae are minor components. The dasyclad Coelosporella jonesi is the most significant contributor. Matrix is similar to associated MA.2.2. Girvanella nodules (up to 1cm in diameter) also occur within this microfacies towards the base of the Trefor Formation (s/m 722, Craignant, Plate 18 fig. B)

10.1.3. Petrography of MA.2.2. Bioclastic packstone to wackestone (Plate 18, fig. E)

This is a petrographically diverse microfacies. The matrix of bioclastic silts and lime/argillaceous muds is often microsparitised or pseudosparitised. Locally significant contributors include crinoids, brachiopods, gastropods, Saccamminopsis sp, textularid foraminiferas and fenestellid bryozoans.

Beresellid and other tubular septate algae, as with Brigantian MA.2.1, are present in very minor proportions. Calcispheres are also minor but ubiquitous components.

Zoophycos and irregular subhorizontal burrows commonly pervade the sediment.

Micritisation of clasts is neglibible, although they commonly show 10 – 20 µm diameter microborings on their surface.

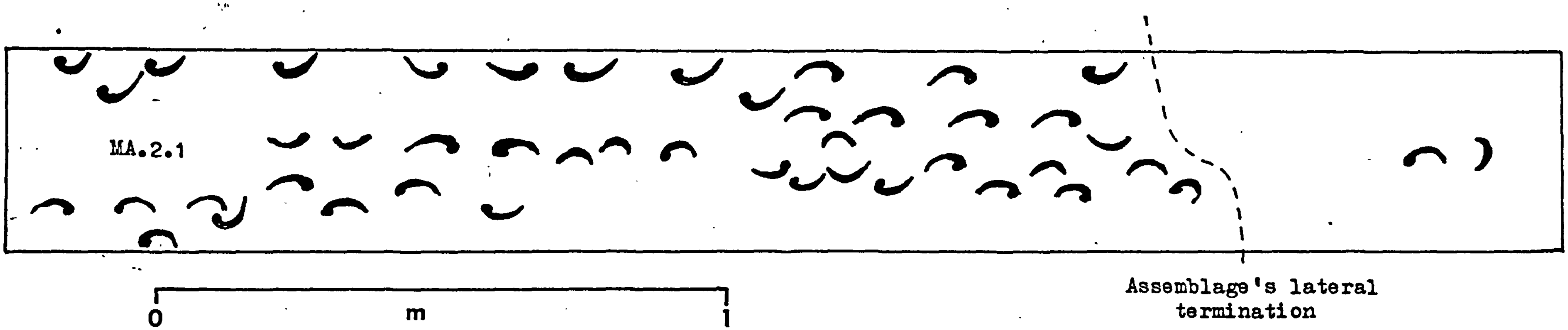
Heterocorals (Hexaphyllia sp) are also rare components of this Brigantian microfacies.

10.1.4. Standardised Petrographic Diversities of MA.2 sediments.

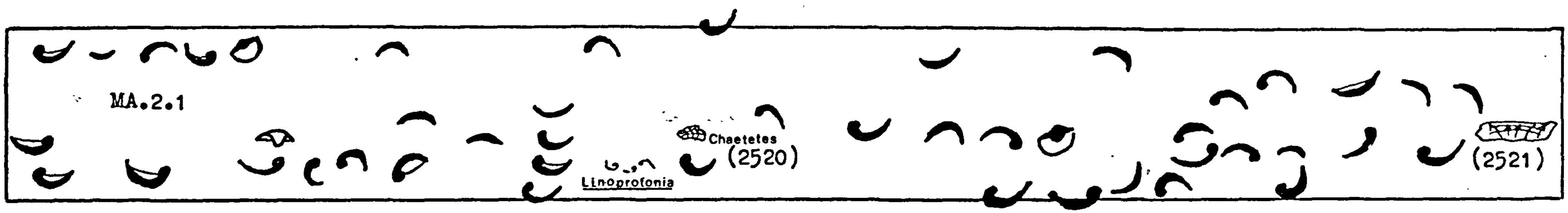
Shown in table 5 , the standardised petrographic diversities of Brigantian MA.2 sediments are the highest recorded in the study-

Lateral log of a Daviesiella bed in the Tynant Formation

A Tynant Ravine (SJ 21964592)



B Eglwyseg Escarpment (SJ 22152638)



SECTIONAL DISTRIBUTION DENSITIES: (S.D.D.)

Defined as the ratio of 'boxed' sectional area of a brachiopod to its average share of quadrat area i.e.

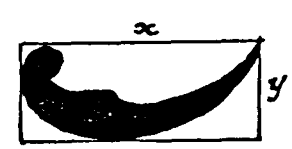
$$\text{S.D.D} = \frac{O}{A.N}$$

Where O = total quadrat area
A = average 'boxed' sectional area
N = number of brachiopods in quadrat area (ventral valves only).

S.D.D for bed 'A' = 5.1

S.D.D for bed 'B' = 4.0

% ventral valves in life orientation for bed 'A' = 44%
" 'B' = 48%



'Boxed' sectional area (A) = x x y

Fig 53

is poor.

7.2.6.3. MICROFLORA

A. Calcispherae. Both Tynant Formation and Sychtyn Member MA.2 microfacies have a diverse assemblage of calcispheres, although they do not dominate the microflora as in MA.1 sediments. Calcispheres from the Tynant MA.2 microfacies include Pachysphaera cf dervillei, Polyderma spp. Vicinesphaera sp. Palaeocancellus cf canaliculatus and Parathurammina suleimanovi. Pachysphaera spp are the most abundant.

B. Red Algae. Ungdarellids are the most common ?red algal group in MA.2 microfacies of both the Tynant Formation and Sychtyn Member. Other red algae include microcolonies of ?Parachaetetes sp. (encrusting form) Stacheoides sp. (Tynant Formation) and Asphaltina sp. (Sychtyn Member). Their distribution is normally sporadic. The ungdarellids have not been observed in growth orientation and are always fragmentary (cf Toomey & Johnson, 1968).

C. Green Algae; Porostromata (Tynant Formation only). Girvanella cf straminea commonly occurs as a loosely woven sediment binding alga (e.g. s/m 006) , and micronodules of Ortonella sp occur (e.g. s/m 016) along sediment laminae.

Dasycladaceae. In both the Tynant Formation and Sychtyn Member Koninckopora sp. is a relatively minor sediment contributor, and does not occur in the volumes encountered in MA.3 microfacies. When present their thalli are fragmentary, and they are often intensely micritised. Coelosporella sp. is also a rare MA.2 dasyclad in the Tynant Formation.

Tribe Palaeoberesellеae. Mamet and Roux. Beresellids are the dominant algal component of MA.2 microfacies, (especially

Kamaenella denbighi) in both the Tynant Formation and Sychtyn Member. Locally beresellids contribute over 60% volume of the rock and over 80% of biota volume. They are orientated bedding parallel, and may have served as a substrate binder, although their presence in ripple laminated (i.e. mobile) sediments casts aspersions on this. Other beresellids are minor sediment contributors (Kamaena cf delicata noteably), along with the problematical ?Uraloporella.

7.2.6.4. MICROFAUNA

A. Foraminifera. MA.2 microfacies have a diverse foraminiferal assemblage, although it rarely accounts for > 5% of the sediment's volume. Endothyrids predominate. In the Tynant Formation these include: Endothyranopsis sp. Plectogyra spp., and ?Pseudoendothyra sp. Other noteable foraminifera are: Ammodiscus spp., Eostaffella sp Palaeotextularia sp and archaediscoids. Significantly the textularids are not as abundant in Lower Asbian sediments as Upper Asbian. Similar foram genera are represented in Sychtyn Member MA.2 microfacies.

B. Ostracods. They are ubiquitous, but minor sediment contributors in MA.2 microfacies, and include both large and small (40μm to 800μm) smooth carapaced and ridge-ornamented forms.

7.2.6.5. PETROGRAPHIC DIVERSITIES Standardised petrographic diversity counts (defined in table 5) have values twice that of MA.1 microfacies association (average 15 ± 4). Each of the three separate groupings within the counts (macrofauna, microfauna and microflora all have values about twice their MA.1 counterparts (5 ± 1.5; 4 ± 1.0; 6.0 ± 2.5 respectively). The high values of macrofaunal diversity (with respect to MA.1) reflect the presence of more stenohaline faunal elements. This diversity, however, is slightly lower than Eglwyseg Formation and Brigantian MA.2 and may imply a slight environmental restriction (see section 7.2.7.1). The high algal diversity values reflect the high algal contribution to the sediment.

7.2.7. Some Aspects of Palaeoecology, Taphonomy and Palaeoenvironment of MA.2

Defining the MA.2 palaeoenvironment requires that it be assessed in comparison with other microfacies associations (especially MA.1 and MA.3) and its stratigraphic context.

It is the basal microfacies association of all cycle forms, and is especially well-developed in the Tynant Formation. It therefore represents the transgressive phase of these minor cycles and grades into either higher energy shoaling (MA.3) microfacies or progradational tidal flat facies (MA.1).

7.2.7.1. SIGNIFICANCE OF BERESELLID DOMINATION. Beresellids have uncertain affinities. They have been classified as simple foraminifera e.g. Loeblich & Tappan (1964) but in more recent works (Elliott, 1970; Petryk & Mamet, 1972; Mamet and Roux, 1974) their dasycladacean affinities have been documented and superficial resemblances to the foraminifera Nodosinella and Moravammina clarified. Riding and Jansa (1974) dismissed Uraloporella (present also within MA.2.1 in both the Tynant Formation and Sychtyn Member) from the Dasycladaceae and reasserted a foraminiferal affinity. Donezella, another remarkably similar branching septate tubular form was considered algal by Rich (1967) but Riding and Jansa (1974) suggested it was also foraminifer, and questioned the algal affinities of all beresellids. Clearly, the morphological similarities, the association, and similar stratigraphic distribution of beresellids, Uraloporella and Donezella implies they are similar types of organism (or are astounding homoeomorphs).

Little has been written on beresellid palaeoecology and growth. Elliott (1970) reconstructed their complete thallus as an erect branching structure. Their cylindrical and radially symetrical pore-wall microstructure supports this contention, and Mamet and Roux (1974, p.140) further reconstruct the genus Kamaenella in a similar orientation. The flat-lying, fragmentary and degraded nature of the MA.2 beresellids indicates some decay and collapse, transportation and fragmentation occurred, further confirmed by ripple laminated sediments. Season (?) blooms may have affected their relative contribution to the sediment and played a significant part in lamination production. Each individual cortex fragment, however, is small enough (normally $\leq$ 200µm) to have been carried in suspension, and deposited as 'winnowed' laminae (e.g. in waning storm stages (Reineck and Singh, 1972)).

The quantities of beresellids present in Tynant Formation and Sychtyn Member MA.2 microfacies implies that local production rates must have been high, and that they probably carpeted the sediment surface as sea-grasses do in Recent environments.

Most extant dasyclads occur in waters < 5m deep of normal marine salinities, and on soft (sand and mud) substrates, in low energy environments, either protected or below normal wave-base (Wray, 1977, p.106). Wray (op.cit.) stresses the problems of applying such known environment tolerances of extant algae to palaeoenvironments of poorly understood extinct algal groups. Nevertheless a broad analogy is feasible. The presence of a relatively diverse algal flora, and a common stenohaline macrofauna including corals, brachiopods and echinoderms, indicates normal marine salinities prevailed. Hughes Clarke & Keij (1973) note that in Recent Arabian Gulf 'lagoonal' areas, echinoderms do not occur in salinities $> 48^{\circ}/_{oo}$ and brachiopods not $> 60^{\circ}/_{oo}$.

Figs. 11 & 12 indicate that relative volume of beresellid material decreases substantially with increased percentage of mud and unidentifiable silt-grade debris, as MA.2 microfacies grade into MA.1. As this apparently reflects a transition to 'prograding tidal-flat' deposition (see section 3.1.) then a shallow subtidal upper limit may logically be placed on their palaeoenvironment range. Mamet and Roux (1974) suggested they grew in lagoonal environments. Riding and Jansa (1974) postulate that Uraloporella (included within this discussion because of it's gross similarity with beresellids) is characteristic of facies 'interior' to carbonate buildups; i.e., in-shore 'protected' shelf. Further to the septate alga/foraminifera analogy, Freeman (1964) and Riding (1970)†, considered Donezella to have a shallow low/energy habitat. (habitat sensu Bathurst (1975, p.102).). Rich (1967) likewise considered both Donezella and Dvinella (beresellid sensu Mamet and Roux (1974)) to be shallow low energy environment indicators, describing them in microfacies remarkably similar to MA.2.1. In contrast, Racz (1964) considered Donezella as tolerant of a moderately turbulent environment (in Bowman 1979, p.33.). The minor quantities of beresellid material included within inter-tidal to supra-tidal MA.1 sediments may be storm-transported sediment load. Their fall-off into MA.3 microfacies is less severe, but is probably due to increased sediment movement, although the relative abundances of hypothetical grazing organisms cannot be overlooked.

Whether beresellid dominatedsubstrates affected the development of macro- and micro-faunal communities is unclear. Bowman (1979) opined that Donezella controlled local ecology "permitting only the smaller biota to inhabit the environment". The diversity of microfauna and flora is relatively high (see table 5). The beresellid carpets probably provided excellent shelter for other

† in Bowman (1979)

algae and benthonic foraminifera.

7.2.7.2. OTHER CONSIDERATIONS. In the Tynant Formation, brachiopod dominated assemblages are 'patchy' in distribution, and of local lateral extent (metres to 100m (i.e. outcrop length)). The MA.2 sediments (especially the more argillaceous ones) do not represent a high energy environment, but large daviesiellids, up to 15cm across their hinge line, are frequently inverted (ventral valve convex up) and their valves separated but have suffered little damage to ornamentation (Davidson, 1874-82). Their close spaced distribution (often within millimetres of each other laterally e.g. fig.53 suggests that this burial assemblage is compositionally similar to the original life assemblage. However, how much both bottom currents and bioturbation were individually responsible for this disturbance is unclear. In Brigantian MA.2 sediments, disturbance of metre-sized coral colonies is common, (section 10. 4. 3). It is significant that in sea-water the relative density of the shell material and high surface area allows disturbance at low bottom current velocities. The dominance of valves of the smaller, but spinose _Linoprotonia_ in life orientation clearly demonstrates the higher stability, provided by its anchor spines. _Linoprotonia_ dominated life assemblages have S.D.D. values approaching unity (e.g. s/m 1041, Plate 11, fig. G).

The whole but disturbed aspect of many coral and sclerosponge colonies further shows that disturbance, with little transport, affected these benthonic organisms.

Calcispheres are abundant, and diverse in MA.2 microfacies, although they do not contribute as much sediment as in most MA.1 lithologies. Their presence implies that either the calcisphere plants grew locally or that they were transported from areas of original production, probably in subtidal MA.1.1 microfacies accumulation zones. A spatial adjacency with MA.1 habitats is, therefore, suggested, as implied by grading contacts between MA.1 and MA.2 in stratigraphic profile.

The lack of micritisation on indigenous clasts suggests that both endolithic and exolithic algal and bacteriological activity was reduced compared with most MA.3 sediments. This is most likely to be a function of reduced water flow through the surficial sediment layer and/or a relatively higher rate of deposition

High insoluble residues of dominantly illitic clay indicates that for the most part bottom currents did not resuspend and winnow MA.2 sediments. This may have been enhanced by beresellid carpets acting as baffles, and algae binding the sediment.

Transgressive events lead to widespread erosion of shorelines. It is probably this erosion of the Lower Palaeozoic siliciclastic hinterland that charged the sediments with detrital clays, and rarer gravel extraclasts (noteable in MA.2.1 and MA.3 microfacies overlying the basal unconformity at Minera G.R. SJ 252523). The depositional setting (see chapter 1) implies that active transgression and erosion was occurring on the Llangollen Embayment flanks (section 2. 5. 1)

7.2.8. Recent Comparative Sediments.

The 'wave-base' concept is appropriately discussed here. Under normal 'fair weather' conditions, Recent sediments are reworked and winnowed by waves down to shallow depths. Under storm conditions, this depth is increased, and coarser grains are both suspended and saltated. The nett result is a sediment of more grainstone and laminated character. Sediments deposited below normal fair-weather wave base, retain their finer matrix.

In a regressive event, sediments deposited below wave base will be superseded by those deposited above. The natural succession of packstone superseded by grainstone fabrics in these Asbian shoaling cycles may reflect this transition from below wave-base (MA.2) to above wave-base (MA.3).

Houbolt† (1957) classified Recent sediments in the Arabian Gulf according to their grain-size fractions and weight percent of insoluble residues, (see composition-field diagram in Bathurst (1975, p.183)). He recognised facies belts (approximately shore-parallel) with increasing insoluble residues and decreasing skeletal sand (his definition : sand of grains < 53μm) towards deeper water (below about 10m and the Iranian shoreline). In the finer sediments insoluble residues vary from 10 to 50 weight %. Clays are dominantly illite and mixed layer illite-smectites. (Facies transitions however are in part controlled by Pleistocene topography producing locally complex facies patterns (Purser & Seibold, 1973)). Wagner & Togt (1973) described these facies as "argillaceous lamellibranch muds" and "lamellibranch muds". Algae are not apparently significant sediment contributors, probably due to the depth of sedimentation.

In lagoonal environments of the Abu Dhabi carbonate complex sediments of mixed-insolubles, lime mud and skeletal sands occur, carpeted by Halodule, a sea-grass, that has important sediment stabilisation effects. This grass, and its algal epiflora help reduce the velocity of bottom-waters (Bathurst, 1975, p.197). Indeed, although strong wind-driven currents cause considerable disturbance of the lagoon water column, "bottom traction of sand is slight."

Hagan and Logan (1974) and Read (1974a) described skeletal packstones and wackestones from the 'basal sheet' of Freycinet Basin and Edel Province, Shark Bay, which have high (25 - 75%)

† In Bathurst, (1975)

algal content, (mostly as a mud and silt degraded fraction). These sediments support a gastropod-codiacean alga community. In this "basin" environment, little wave reworking occurs, and obliteration of primary structure by bioturbation is intensive (Read, 1974a, p.30).

7.2.9. Fossil Comparative Sediments.

The Lower Asbian of Ravenstonedale (Potts Beck Limestone) has similar MA.2 lithologies developed within similar minor cycle forms (see chapter 5) and which have high beresellid algal contributions.

The high argillaceous content of MA.2 microfacies gives them superficial resemblance to Lower Carboniferous 'basinal' sediments, i.e., thinly bedded argillaceous wackestones, and calcareous shales. In these sediments, bioclastic material is scant, and much of it derived from shelf areas by turbidity transport.

MA.2 lithologies are very common throughout the Asbian and Brigantian of North Wales and the other 'shelf' areas of Britain.

Bowman (1979) described "*Donezella* bafflestone facies" from the Middle Carboniferous of the Cantabrians, NW Spain. These bafflestones form lensoid 'mounds', with 10-25% by volume of *Donezella*. Intermound bioclastic packstones and wackestones also contain *Donezella* along with fragmentary brachiopods, crinoids and foraminifera. Significantly these facies occur in a cyclic sequence and are developed at periods of "maximum transgression". The similarity between these sediments and the Tynant Formation (in particular) indicates the cosmopolitan importance of these septate tubular alga/foraminifera as sediment contributors, and partly as sediment modifiers.

7.3. ALGA PELOID GRAINSTONE MICROFACIES ASSOCIATION (MA.3)

7.3.1. Distribution and Field Character of MA.3

MA.3 microfacies are dominant lithologies within the Sychtyn Member cycles and upper cycles of the Tynant Formation. They include the pale weathering, light grey non-argillaceous bioclastic limestones of Somerville (1977).

MA.3 microfacies usually overlie MA.2 and underlie MA.1 within each cycle, and may grade from one to the other or have sharp bedding plane transitions.

Bed morphologies vary between thin lensoid horizons only a few cm thick, to beds $\sim$2m thick. The very thick (> 2m) uniform MA.3.1 beds of the Eglwyseg (and Llynclys) Formation occur in neither the Tynant Formation nor the Sychtyn Member.

Whereas MA.3 lithologies characterise cycle tops in the Eglwyseg and Llynclys Formations, they are normally internal or basal to cycles in the Tynant Formation and Sychtyn Member respectively.

Sediments are paler to dark grey, often obviously bioclastic in the field, but vary between coarse 'biosparites' and fine-grained 'biomicrites' with minimal insoluble residues. They include the 'clotted pelsparites' (structure grumeleuse of Cayeux (1935)) of Bathurst (1975, p.86), and the typical mottled 'pseudobreccias', although both features are more pervasive in Eglwyseg Formation-type facies.

PETROGRAPHY OF MA.3 SEDIMENTS

7.3.2. Petrography of MA.3.1. Alga Peloid grainstone.

MA.3.1 is the most common MA.3 microfacies. In the Sychtyn

PLATE 12

A Low angle cross laminae within alga-peloid grainstone, Tynant Formation. G.R. SJ 22144633, Llwyn Hen Parc.

B Peloid Grainstone MA.3.3 s/m 1074, Sychtyn Member, G.R. SJ 25202500

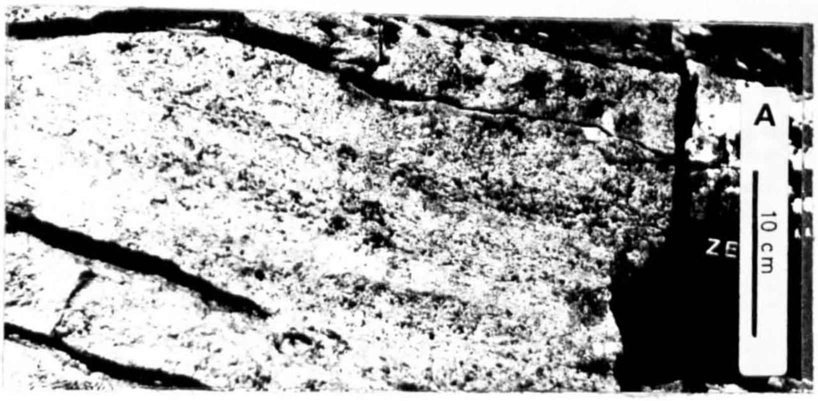
C Alga Peloid Grainstone MA.3.1 (poorly sorted) s/m 1377. Sychtyn Member, Craig y Rhiw, G. R. SJ 23802975.

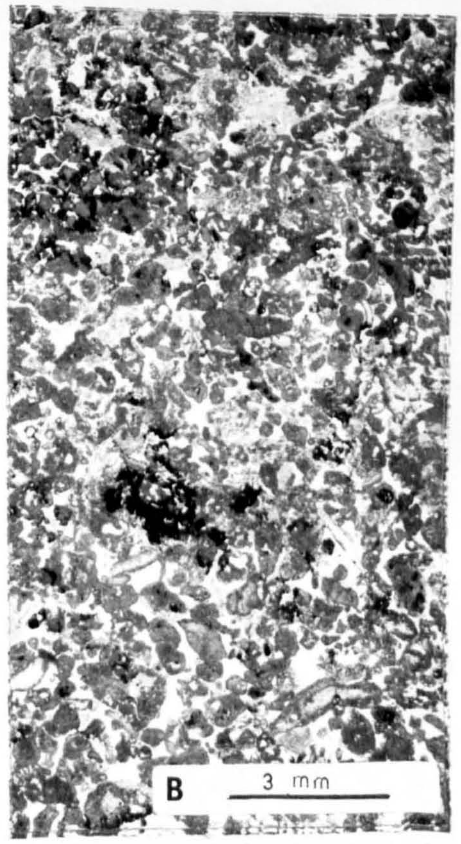
D Alga Peloid Grainstone MA.3.1. (fine sands). s/m 194. Tynant Formation. G. R. SJ 23424947 (flanks of Cyrn y Brain) (Negative print).

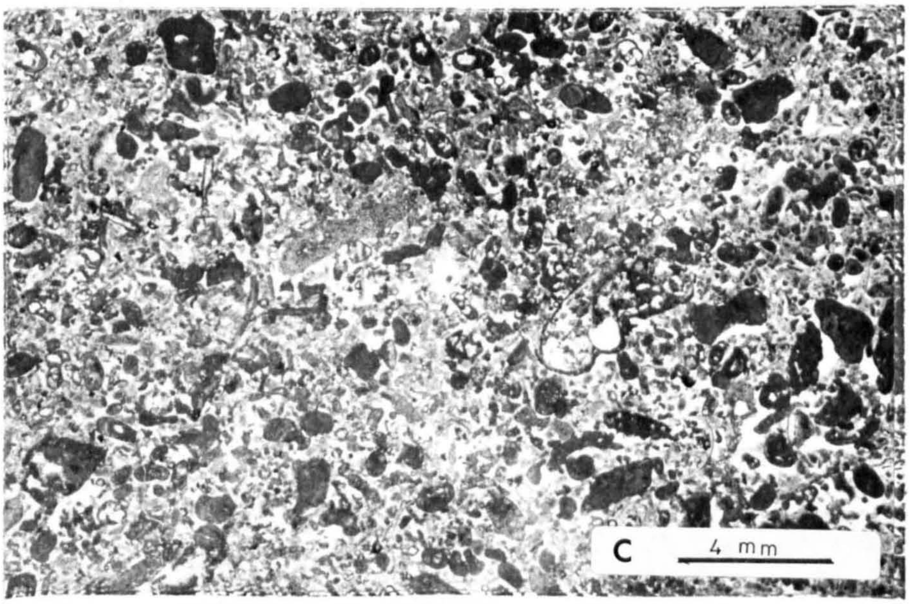
E Oncolitic growth around gastropod clast. s/m 193, Tynant Formation. Transgressive phase of a Minor cycle. Significant is the presence of incorporated clasts and fenestral fabrics within the overgrowth. Calcareous (Garwoodia-form) filaments are rare G.R. SJ 23424947 (flanks of Cyrn y Brain) (see Frontispiece)

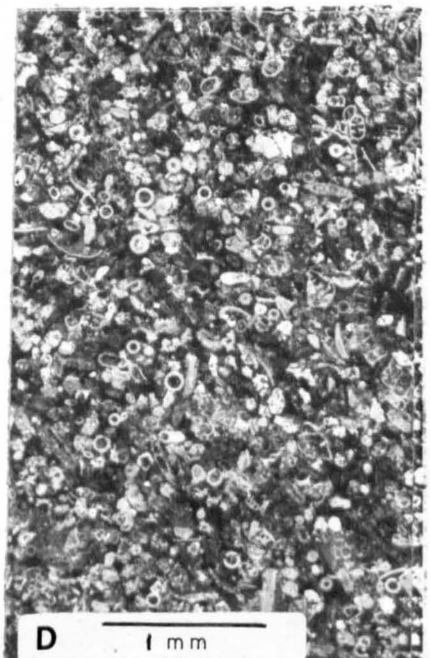
F Burrow structure within fine MA.3.1 sands s/m 140 Tynant Formation, World's End, G. R. SJ 23304788. (Negative print from peel).

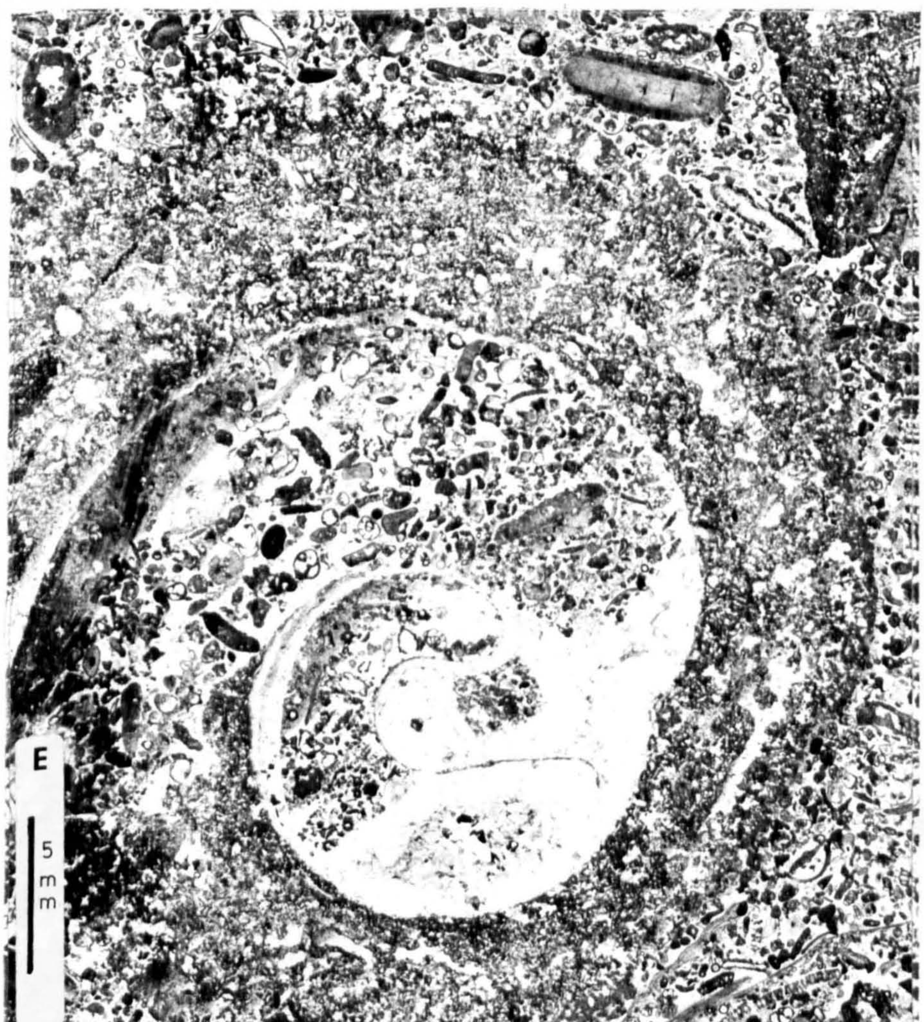
PLATE 12

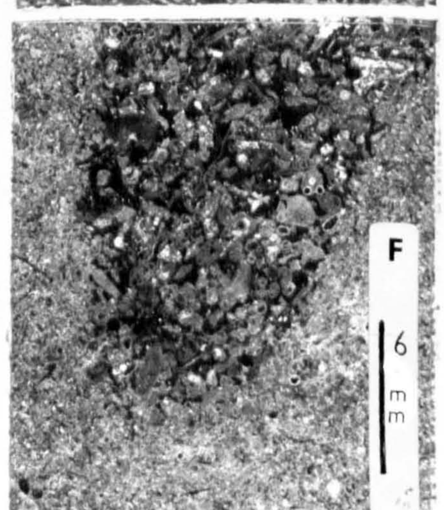












Member it is common to all cycles. Both parallel and low angle cross lamination on a centimetre scale is variably developed, and when present depends on variations in clast types, clast sizes or texture. Much of the sediment has an homogeneous fabric, but large cross laminated sets (20cm bedform relief) do occur e.g. G.R. SJ 23602955 (Sychtyn Member) and G.R. SJ22144633 (Tynant Formation) (Plate 12, fig. A), and parallel laminated sediments are commonplace. Mottling does not occur on the pervasive scale of the Eglwyseg Formation. Large (5cm to 15cm long axis) laterally elongate, sharply defined and smooth margined mottles of darker sediment in lighter do occur in Sychtyn Member sediments (see chart C) (cf. Plate 17, figB). The darker mottles have a more prominent grain-support texture, and are compositionally similar to the surrounding sediment. In extreme cases these mottles are defined by stylolites.

Characteristically, this microfacies contains a heterogeneous mixture of variable sized peloids and intraclasts (10mm to 20µm), micritised and non-micritised bioclasts, with both endolithic and exolithic coatings, including a diverse assemblage of algae and foraminifera (Appendix IV). In most cases the matrix comprises calcite spar and crystal silt grade 'cement', together often giving the appearance of '<u>structure grumeleuse</u>'. General size range and 'sorting' varies, from finely bioclastic calcisphere-peloid-foram grainstones that seldom have clasts exceeding 300µm (Plate 12 fig. D) to the irregular hotch-potch of clasts ranging up to centimetres in size, (Plate 12 fig. C) in coral-brachiopod burial assemblage sediments, more typical of Eglwyseg Formation MA.3.1.

Algal contributions are high in many of these sediments, from both the Tynant Formation and Sychtyn Member. Tubular and septate ?algae/beresellids account for up to 74% by volume of the sediment present (s/m 1374 Sychtyn Member (mean 38% ± 20% for Sychtyn Member)). The dasyclad Koninkopora spp is ubiquitous to MA.3 lithologies, in greater volumes than MA.2 lithologies. MA.2.1 microfacies grade into MA.3.1 with grain-stone packing, and reduction of lime mud and silt. Echinoderm debris is the dominant macrofauna contributor, although it rarely exceeds 5% by volume. In burial assemblage sediments, the dominant macrofauna contributors reflect the whole-fossil assemblage.

Within c.4m of the basal angular unconformity north of Worlds End MA.3.1 microfacies contain well developed nonskeletal and skeletal oncolites,(e.g. s/m 192) both coating clasts and as irregular nodules up to 5cm diameter. (Plate 12,fig. D). Algal coatings are Garwoodia sp., and?Polymorphocodium sp. occurring together within individual oncolites. Garwoodia sp also occurs in association as micronodules. These proximal-to-unconformity sediments also contain Lower Palaezoic extraclasts (up to c.10% by volume) which become less common at stratigraphically higher levels, although at Minera they occur in the basal 5m of the overlying Eglwyseg Formation. In this microfacies development, beresellid algae are rare.

Floatstone textures (sensu Dunham, 1962) are included within MA.3 where matrix is MA.3, but where larger (> 2mm) clasts provide a framework support to the sediment it is here classified as MA.4. (Bioclast Peloid Rudstone Microfacies association)

Rarely bioturbation is visible as irregular and tubular (c. 1cm diam.) zones of coarser, less compacted sediment, at all orientations to the primary sediment lamination (e.g. s/m 140, and 1461; Plate 12 fig. G).

Compaction is visible in these sediments on two scales:

Large scale microstylolitic compaction, that retains little cement matrix, (e.g. s/m. 195) and smaller scale compaction, shown by compaction-fracture of e.g. ostracods and chaetetid colonies.

7.3.3. Petrography of MA.3.2. Bioclast grainstone

With decreased volume of peloids and beresellids MA.3.1 grades into MA.3.2. Typically, MA.3.2 is a lime sand composed of a significant (> 20% of total volume) proportion of recognisable (not micritised peloids) bioclasts of macrofossils. They are the 'grainstone' equivalent of MA.2.2. This microfacies, therefore, includes a diverse suite of compositions (see Appendix 11) of "immature" sediments (see section 9. 3. 1). Foraminiferal microcoquinas (terminology of Ball, 1971) are not included within MA.3.2, although they are rare elements of the Tynant and Sychtyn sediments.

In the Tynant Formation and Sychtyn Member MA.3.2. microfacies are rare. They occur as laterally impersistent (traceable over 100's of metres) beds, up to 80cm thick (see chart A), and in the Tynant Formation of the Llwyn Hen Park area appear as coarse bioclastic deposits in marked contrast to the surrounding MA.2 and MA.1 microfacies. The clasts are diverse, including much fragmentary coral material, but containing neither whole, nor large clasts of coral colonies (e.g. s/m. 473). Vague lamination is usually present, parallel and on a centimetre scale.

7.3.4. Petrography of MA.3.3 Peloid grainstone

Another rare microfacies in the Tynant and Sychtyn Formation, this represents the opposite end member on the 'maturity' spectrum (see section 9.3.1.1) to MA.3.2. They are defined as containing 50% by volume of peloids or intraclasts. As with other

MA.3 lithologies, clast sizes are variable, and with increasing clast sizes MA.3.3 grades into rudstones or intraformational conglomerates. Compaction phenomena are similar to MA.3.1..

7.3.5. Early Diagenetic features associated with MA.3

7.3.5.1. MICRITIC RIM/MENISCUS CEMENTS Within MA.3 microfacies, especially in the Minera area, within the Tynant Formation, 5 to 150μm isopachous rims of dense dark brown micrite, both coat and rarely have meniscus thickenings between adjacent clasts (Plate 13 figs.D & B) (e.g. s/m 092 from 20cm beneath a palaeokarstic surface, Tynant Formation, Minera). The meniscus effects are exceedingly pronounced between some clasts, where 5 to 20μm impersistent rims broaden to > 200μm thick. The lack of lamination in the cement does not allow observation of growth trends at the meniscus contacts. Besides meniscus thickenings, the micrite locally thickens into cavities above and below clast surfaces, but in a non-uniform manner.(Calcrete linings are associated with primary fossil voids, diminishing in extent further from the fossil outer surface e.g.s/m 092, similar to micrite described in section 11.3.1.3). In specimen 092, this meniscus micrite has a fringe of fine acicular cement (see next section), overlain by a further micrite coating.

In section 6.4.10 , similar cements (not displaying a meniscus effect, but within MA.1 where grainstone texture is rare) were argued as having a probable beach rock origin. A supratidal origin is suggested here by their association with palaeokarstic surfaces and meniscus phenomena that according to Dunham (1971) reflects precipitation in the vadose zone, in which porewaters are mixed with entrapped air.

7.3.5.2. ISOPACHOUS ACICULAR CEMENTS. Within MA.3 microfacies, this is a common early cement type, found either in association with micrite cements or as the only rim cement. Where associated with micrite cements an outer micrite coat faithfully preserves fine detail of the acicular terminations, although the crystals are degraded to equant microspar (Plate 13 fig. C). They are similar to acicular cements described in section 6.4.9, but no bladed terminations have been identified.

In places, hour-glass meniscus effects (cf. Tucker, 1975) within the acicular/microspar fringe of s/m092 suggest entrapped air pockets (Plate 13, fig C), and, therefore, a vadose cement origin.

The rims are 50 to 300μm thick, individual crystals rarely greater than 30μm broad, and are difficult to see as a distinct cement generation, when overlain by spar druse growing centripetally away from this cement surface. This cement fabric resembles many recent beachrocks and hardground cements (Dravis, 1979, Fursich, 1979), but its association with meniscus ?calcrete micrite cement suggests formation associated with a fluctuating water table (causing alternate vadose/phreatic cementation). Whether precipitation was from marine or meteoric waters is unclear. The fringe resembles some Recent beach rock cements (Stoddart and Cann 1969)

7.3.6. FOSSIL ASSOCIATION AND DIVERSITY IN MA.3.

7.3.6.1. MACROFAUNA Macrofaunal assemblages and distributions within Tynant Formation and Sychtyn Member MA.3 microfacies are very similar to those described in section 3.2.5 for MA.2 microfacies. In the Tynant Formation Daviesiella llangollensis dominated assemblages occur infrequently, along with scattered chaetetid colonies, productids, and corals. They are invariably not in growth orientation, and have a degree of fragmentation and abrasion.

In the Sychtyn Member syringoporoid-chaetetid biostromes are common, but their densities rarely exceed 1 colony/m over 20m of outcrop. Rugosan corals are very rare in MA.3 microfacies of the Sychtyn Member.

Athyrids (Composita sp.) are prolific in discontinuous biostromes. In these sediments, the brachiopod valve prisms contribute, as individual grains, significant proportions of the sediment. Linoprotonia occurs infrequently in comparison to MA.2 microfacies, and does not form the in-growth-orientation assemblages characteristic of MA.2.

7.3.6.2. MICROFLORA Beresellid algae (mostly Kamaenella denbighi) are the most widespread and prolific, but are not as ubiquitously dominant as in MA.2.1. Kamaena cf delicata becomes a common beresellid in MA.3. They occur along with Uraloporella. Ungdarellids are common in MA.3, but Stacheoides has not been recognised.

Koninckopora becomes a locally dominant algal element and is ubiquitous. Calcispheres are very variable in distribution. In finer bioclastic MA.3 lithologies they may be very abundant, accounting for up to 6% by volume, but mostly < 4%. Pachysphaera is the most common genus, but overall the genera are as diverse as in MA.2 microfacies.

7.3.6.3. MICROFAUNA. MA.3 sediments also contain a diverse foraminiferal microfauna with similar forms to MA.2 sediments, but notably many are micritised and have micrite chamber fillings, in extreme cases becoming (un)recognisable peloids.

7.3.7. Some Aspects of Palaeoenvironment and Palaeoecology of MA.3

7.3.7.1. SEDIMENT MOVEMENT The grain-support and matrix-poor nature of MA.3 microfacies suggests that deposition of lime-mud and silt fines was minimal due to either rapid deposition and a slow rate of production of the fines, or that bottom currents did not permit fines to settle.

The common presence of lamination suggests that deposition was controlled by rhythmic events, and the mixture of micritised and unmicritised grains suggests that either clasts of different histories and generations were admixed, or that sedimentation was slow enough to allow earlier deposited clasts to become algally micritised, (inferred from Kendall and Skipwith (1969, p.850)) whilst others (of a similar type) escaped the process. This latter unfavourable model leads to the support of the "transport-mixing" model (see section 9.3. 1). Cross laminated sediments attest to moving bedforms, and lack of fines in the more massive MA.3 deposits suggests deposition probably above fair-weather wave-base.

The recognition of shelf storm deposits within beds of MA.3 microfacies (section 7. 4. 3) attests to the potentially high degree of sediment movement.

7.3.7.2. DIVERSITIES. Petrographic diversity measurements of Syohtyn and Tynant MA.3 microfacies (table 5) reveal values very similar to MA.2 microfacies (total standardised petrographic diversity 16 ± 3). Algal diversities are similar to MA.2 (7 ± 1), as are macrofauna and microfauna. (4 ± 1 and 5 ± 2 respectively). The relatively impoverished macrofaunal assemblage (impoverished against the potential elements available) imples that although stenohaline faunal elements are relatively abundant (<u>cf.</u> MA.1) there was a degree of environmental restriction (definition of Purser and Seibold 1973, p.7) In the palaeogeographic situation envisaged here, of a broad shallow and embayed carbonate shelf (section 1. 3. 3), normal oceanic salinities are less likely, and metahaline conditions (possibly even locally hypersaline) probably existed. Much of the shallower shelf region of the Arabian Gulf has a salinity $> 40^{o}/_{oo}$, but supports a diverse stenohaline fauna, (Purser and Seibold, 1973)

7.3.7.3. OTHER CONSIDERATIONS. Tubular septate algae (beresellids and?<u>Uraloporella</u> dominantly) are principle microfloral elements in these microfacies, as in MA.2 microfacies. Their fragmentary state, and variable degrees of medular and cortical micritisation indicate transportation (possibly as lithoskels) and degradation occurred, implying that in MA.3 habitats they did not have a sufficient sediment stabilisation effect to counter the bottom-sediment movement.

7.3.8. <u>Comparative Sediments</u>.

7.3.8.1. RECENT. Many recent low latitude carbonate shelf environments are actively producing MA.3 type carbonate sands.

On the Bahama Bank the "Grapestone Lithofacies" is an heterogeneous deposit of skeletal and peloid grains, but is partly cemented by micrite into grain aggregates. These are the 'stable' sands of Bathurst (1975, p.121), stabilised by both early cementation and subtidal algal mat coverings. They occupy a broad central-marginal belt to the Bank, and are internal to the narrow marginal unstable oolitic and coralgal sands which may include > 50% peloids (Purdy, 1963) and > 10% calcareous algae. The stable sands are only moved during the highest tides and storms, and it is probably the lack of particle movement that induces the early micritic cementation and grapestone formation. Kendall and Skipwith (1969, p.866) describe 'aggregate ' sediments forming in "moderately exposed" lagoon shores and offshore banks which are partly protected from "intense hydrodynamic activity". The lack of grapestone cements in the Tynant and Sychtyn MA.3 sediments adds further weight to the transport-mixing model.

The unstable sands are ripple marked (due to vigorous tidal currents) (Bathurst, 1975, p.116), and in the extreme have dune bedforms up to 3m relief. The stable sands, however, (occurring in waters < 9m, and occasionally intertidally emergent, have less common 'ripple marking' (Newell _et al_., 1959, p.220) although apparently stable megaripples occur within tidal channels transecting this belt.

MA.3 microfacies occasionally do have bedform relief, showing that sediment transport was locally significant.

Similar sediments also occur in shallow sublittoral environments: (1) in the Gulf of Batabano, Cuba, also in central-marginal shelf areas, as 'ovoid grain sand' and 'composite grain sand' (Daetwyler and Kidwell, 1959, in Bathurst, 1975, p.172) in normal salinity waters. _c._ 10m deep; (2) In the Shark Bay complex, as intertidal veneer pellet grainstones (with lithoclasts and lithoskels locally abundant), as unstabilised sublittoral sheet 'lithotopes' Read, 1974, p.37), where wave action continuously reworks the sediment, and in the sublittoral bank lithotope as skeletal lithoskel grainstones, with common compositional lamination; (3) In the Arabian Gulf (Wagner and Togt, 1974) as molluscan sands and muddy sands in waters <40m deep.

7.3.8.2. ANCIENT. MA.3 microfacies include the phase 'C biopelsparites of Sommerville (1977, 1979a), in overlying late Asbian minor cycles (see on). MA.3 microfacies occur in similar cycle forms within the Potts Beck Limestone (see section 5. 1), and is the dominant microfacies association of Asbian sediments on the Askrigg (Kingsdale Limestone) and Derbyshire 'Blocks'.

7.4. INTRAFORMATIONAL CONGLOMERATE LITHOFACIES (L.7) AND BIOCLAST PELOID RUDSTONE MICROFACIES ASSOCIATION (MA.4)

7.4.1. Distribution and Field Character of L.7.

MA.4 & L.7 facies are rare elements of Tynant and Sychtyn sediments. They characteristically occur at the base of minor cycles, within MA.3 and within MA.1 microfacies. Some are laterally persistent across 100's of metres of outcrop, whilst others are lensoid deposits only a few metres across. They rarely exceed 50cm thickness. Clast types depend upon underlying (and adjacent) sediments, but include a full spectrum of lithologies and fossil materials.

PETROGRAPHY OF L.7&8 SEDIMENTS (See section 6. 2. 6)

7.4.2. Petrography of matrix supported conglomerates.

This is a widespread and variable lithology. In the Tynant Formation, it occurs characteristically on cycle bases, overlying scoured or 'sutureddiscontinuity' surfaces (section 11. 1) comprising clasts of underlying MA.1 lithologies (and rare extraclasts) within the transgressive MA.2 or MA.3 sediments .

At G.R. SJ 22034535 (Tynant Ravine) a coarse heterolithic intraclastic conglomerate (clasts up to 20cm diameter) overlies a cycle boundary, with MA.2.2 matrix. (Plate 1, fig. B). Less exceptional deposits have well-separated clasts that decrease in abundance rapidly above and away from the cycle boundary.

(Extraformational conglomerates, common in the Tynant Formation of the Minera area are of L.8.2 form, occurring both within MA.2 and MA.3 matrices (see chart A). None have been observed clast supported.)

7.4.3. Petrography of MA.4.1 Bioclast peloid rudstone.(Plate 13, fig A)

This is also a minor microfacies in the Sychtyn Member and Tynant Formation. It includes coquinas and peloid rudstones in which MA.3 lithologies may be present as matrix, but where clasts > 2mm provide a framework support to the sediment. Where clasts, bioclasts and peloids > 2mm do not provide a framework support texture to the sediments, they are included as 'Float-stones' within the microfacies of the matrix.

In the Sychtyn Member, brachiopod and gastropod coquinas occur as lensoid to laterally persistent layers, within MA.3 lithologies, and up to 20cm thick, with MA.3.1 matrix to the whole and fragmentary bioclast debris. They have vague to well-defined centimetre lamination, and include a suite of 'multigeneration' clasts and often diverse clast types (note that indigenous faunal elements are normally predominant, e.g. s/m 2150).

Rudstone microfacies have many large intergranular 'voids', caused both by umbrella and loose packing features.

7.4.4. Some Aspects of Fossil Associations within L.7 and MA.4

MA.4 sediments, apparently by their rapidly deposited and

PLATE 13

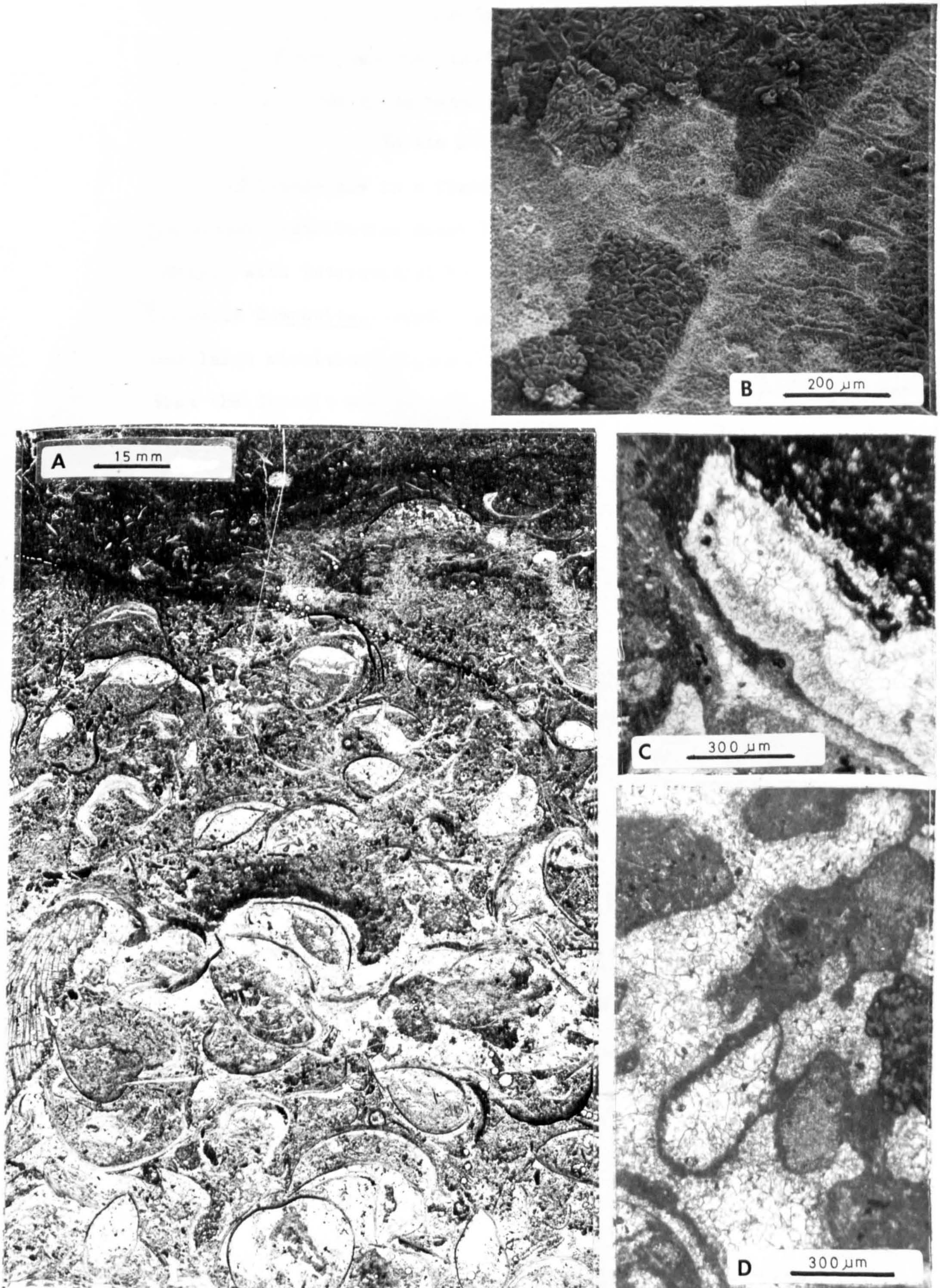
A A 'Storm Bed' of Composita sp; mixed assemblage of clasts; Chaetetes sp; fragmentary and complete Composita. Note geopetal features and lack of sediment between many adjacent brachiopod clasts, suggesting rapid deposition S/m 2150, Sychtyn Member G. R. SJ 25232503. (MA. 4.1)

B S.E.M. photomicrograph of meniscous micrite between clast of Koninckopora and peloid. S/m 092. Tynant Formation, Minera. G. R. SJ 25265226.

C Vadose meniscous 'hour-glass' structure of microspar cement overlying an acicular microspar cement fringe. S/m 092. Tynant Formation, Minera

D Dense (pale) micrite cement rind with meniscous structure as shown in fig. B. S/m 092.

PLATE 13



hydrodynamically severe depositional mechanism, provide biased evidence of fossil communities. Locally transported and *en masse* deposited coquinas (MA.4.1) reflect living assemblages. S/m 2150 is cited as an example from the Sychtyn Member (Plate 13 fig. A). In the most dense zone of the fossil band, the brachiopods are in a framework support (MA.4.1) texture (with sectional distribution densities (S.D.D; see fig. 53) less than unity.) with interstitial MA.3.1 sediment. The association of ?juvenile *Composita*, micritised and non-micritised shell fragments, and large micritised fragments of chaetetid colonies, suggests that the deposit was formed by admixing of more 'mature' (see section 9. 3. 1) sediment, with a life assemblage ⟹ a mixed death assemblage (Raup and Stanley, 1978). The large proportion of void (now infilled by drusy spar) interstitial to the larger bioclasts (as well as much umbrella and fossil internal-cavity) implies a rapid deposition, precluding the gradual settlement and better sediment packing expected of more slowly deposited sediment.

L.7.2 sediments support faunas of similar assemblage-types to MA.3 sediments. In the Tynant Formation *Daviesiella* shell beds are associated with L.7.2 sediments.

7.4.5. Palaeoenvironments of L.7.2 and MA.4.1.

The poorly sorted nature, overall distribution and high current velocity required to transport the clasts of MA.4 microfacies suggests that abnormal conditions were responsible for their formation. Plausible mechanisms are either abnormally high tidal-currents or meteorological storms. These mechanisms have been documented in recent environments, and their ensuing deposits investigated, but more information is available on siliciclastic than carbonate environments.

7.4.5.1 MATRIX SUPPORTED CONGLOMERATES. Occurring on cycle bases above planar and S.D. surfaces, these deposits represent the basal erosive event of the transgressive shore-zone. Their matrix-supported form (MA.3 & MA.2 microfacies matrices) implies that intraclasts were admixed with actively accreting sediment, by transport from an adjacent shore 'platform'. The rapid decrease in clasts (size and quantity) above the cycle base may reflect the rapidity of the transgressive event, as active shore zone erosion moves further 'landward', and active deposition increases further offshore. The irregularities in sutured discontinuity surfaces may have provided sediment traps even intertidally (as 'rock-pools' - see section 11. 1. 2) for locally eroded intraclasts. Kasig (1980) has described similar intraclast rich sediments (and 'sedimentary breccias') from V2b Dinantian cycle bases of the Aachen region.

7.4.5.2. BIOCLASTIC PELOIDAL RUDSTONES. The mixed generation and type-variability of the clasts, very open packing, lack of sorting, and bed thinness, are evidence for rapid deposition. Storms or abnormally high tidal currents are believed responsible (the latter is unlikely - see chapter 6). This type of storm deposit corresponds to the 'parautochthonous coquinas' of Aigner <u>et al</u>. (1979) (See section 7. 4. 3).

Many siliciclastic shelf storm deposits are characterised by a pseudoturbidite structure (e.g. Brenchley <u>et al</u>. 1979), with upper parts of the single-event deposits representing settlement of fines from a suspension-charged water column in the storm's waning stages. This, however, is controlled by water-depth, storm strength and direction, and sediment availability. In these Asbian sediments this latter stage of storm deposition will not be readily recognised from the background MA.3 sediments with which they are usually associated, especially after bioturbation and a degree of surficial sediment movement has occurred once the storm has passed. High velocity bottom currents (e.g. the storm surge ebb) would readily transport the coarser sediment fraction.

7.5. SUMMARY OF SUBTIDAL MICROFACIES ASSOCIATIONS.

7.5.1. Alga Packstone Microfacies Association

Alga packstone microfacies association (MA.2) are basal lithologies to minor cycles, and are most common in the Tynant Formation. Characteristically, they are thin bedded dark argillaceous units, but have a more diverse stenohaline faunal assemblage than MA.1. The beresellid alga Kamaenella denbighi is the dominant bioclast, locally accounting for over 60% by volume of sediment. Clasts are seldom micritised.

Two microfacies are recognised: beresellid wackestone/packstone (MA.2.1) and calcareous shale (MA.2.3). In both, green and red algae, and foraminifera are diverse. Compaction enhances packstone textures, and fragments many clasts. Microstylolites are pervasive, and both parallel and cross-laminae common.

Both tabular and nodular cherts occur, with a distinct dolomitic zonation structure peritheral to the nodules. Evidence of their early formation is shown by silica cement growth; bioturbation disruption (?); pre-stylolitisation origin; and retention of opaline silica spicules before active dissolution had removed them. The mixing zone model proposed by Knauth (1979) is accepted here.

Early acicular cements are rare and restricted to primary bioclast cavities.

Fasciculate lithostrotiontids, syringoporoids and chaetetids are dominant non-brachiopodal macrofaunal elements, especially in the Sychtyn Member. In the Tynant Formation, Daviesiella llangollensis and Linoprotonia hemisphaerica are dominant elements of brachiopod assemblages. The ventral spines of the productid increase its bottom stability with respect to the much larger and robust chonetids that are mostly overturned from life orientation. In the Sychtyn Member, Linoprotonia, and Composita sp. dominate brachiopod assemblages.

Standardised petrographic diversities are substantially higher than MA.1 microfacies (mean 15 ± 4), being approximately 2x higher in both macrofauna and microfauna and flora.

Beresellid and septate (?) algae may have acted as sediment carpets, helping to trap and bind sediment particles, sheltering them from current activity, although their fragmentation and transportation is locally extreme

Their presence as transgressive basal phases of minor cycles and presence of a 'finer' matrix, suggests deposition below fair weather wave base.

Comparison with Recent environments provide analogues with lagoonal situations of the Abu Dhabi complex, sediments from the 'basal sheets' of Shark Bay, and the offshore muddy sands of the Arabian Gulf.

7.5.2. Alga Peloid Grainstone Microfacies Association.

This microfacies association (MA.3) includes poorly sorted alga-peloid grainstone (MA.3.1); bioclast grainstone (MA.3.2.); and peloid grainstones (MA.3.3). They are dominant lithologies in the Sychtyn Member and occur within mid-zones of Tynant

Formation cycles. Towards the Minera area, in upper cycles of the Tynant Formation, MA.3 microfacies replace MA.2 in dominance. They are not developed to the extent of the Eglwyseg Formation MA.3.

MA.3.1, MA.3.2, MA.3.3, represent a spectrum of compositions. MA.3.2 is characterised with > 20% by volume of recognisable bioclasts, and MA.3.3 with > 50% peloids. MA.3.1 encompasses other compositions. Clast type and size-sorting is very variable, from very finely clastic (≤ 300μm), to MA.4 rudstones (≥ 2mm).

Beresellids are also dominant algal components in MA.3, along with other tubular-septate ?algae. Koninkopora spp is more abundant in this microfacies association than MA.2, and crinoidal bioclasts are the dominant macrofaunal contributor. Skeletal oncolites occur in Tynant Formation MA.3 sediments. The foraminiferal assemblage is similar to MA.2. Locally fragmentary macrofaunal associations modify the sediment composition (from MA.3.1 to MA.3.2).

The sediment transport mixing model as described in section 9. 3. 1 applies to these sediments.

Early cement fabrics are rare, and are mainly developed in MA.3 lithologies from the Minera area Tynant Formation. They comprise micritic meniscus, and isopachous acicular cements. A beach rock origin is suggested.

Faunal associations are similar to those of MA.2 microfacies, as reflected in their similar petrographic diversities (16 ± 3 for MA.3). Chaetetid-syringoporoid biostromes are notably common in the Sychtyn Member.

MA.3 microfacies are comparable with many Recent bioclastic and peliodal lime sands, including the "Grapestone Lithofacies" of the Bahama Bank, shallow sublittoral sands of the Gulf of Batabano, and both intertidal veneers and sublittoral sheet lithotopes of the Shark Bay complex. They represent sediments deposited in shallow, sometimes shoaling, subtidal environments, with near-oceanic salinities, and supporting a relatively diverse benthos locally. Sediment movement is common, rarely producing high-relief bedforms, and accounts for the mixed generations of bioclasts and peloids. Storm deposits (of MA.4) are associated.

7.5.3. Intraformational Conglomerate and Bioclast Peloid Rudstone.

Other facies associated with the subtidal sediments are matrix supported (L.7.2) conglomerates (mostly of intraformational origin) and bioclast-peloid rudstones, associated with MA.3 microfacies.

L.7.2 are more common as transgressive deposits on cycle bases and are products of the initial bedrock erosion.

MA.4.1 sediments are associated with MA.3 sediments in thin, laterally impersistent units. They include coquinas (especially Composita dominated, in the Sychtyn Member) that are interpreted as storm deposits, reflecting local life assemblages.

L.8.2 sediments reflect erosion (?associated with the transgression) of an adjacent Lower Palaeozoic hinterland: specifically the axial region of the Cyrn y Brain on the northern margin of the Llangollen embayment.

8. SEDIMENTS OF THE LLANYMYNECH MEMBER

Section 2.4. reviews the stratigraphy and cycle form of this member. As with the succeeding microfacies analysis chapters, repetition of information and interpretation described and discussed in chapters 6 and 7 will be avoided where possible.

Microfacies associations MA.1; MA.2; MA.3; and MA.4, are represented in the Member, and discussed under these headings.

Microfacies described here, and not in previous chapters are:

Alga Packstone Microfacies Association (MA.2)

MA.2.2. Bioclastic packstone

Alga Peloid Grainstone Microfacies Association (MA.3)

MA.3.4. Oolitic grainstone

8.1. PETROGRAPHY OF LLANYMYNECH MEMBER SEDIMENTS.

8.1.1. Calcisphere wackestone microfacies association. (MA.1)

MA.1 is the least common microfacies association, accounting for about 1% volume of the member, occurring as three $\leqslant$ 30cm horizons at Llanymynech (chart C), marking cycle tops (?), one with an associated sutured discontinuity surface† (see section 11.1). The sediments grade upwards from MA.3 microfacies as in Sychtyn Member cycles, and rarely have small irregular, tubular, and packing fenestral fabrics (section 6.4.2). Internal sediments, (section 6.4.6) and 'ped' clots (section 6.4.12.2) are variably developed. Most are MA.1.1 (calcisphere mudstone to packstone), some with vague mm lamination, but MA.1.4 occurs (e.g. s/m 1406).

At G.R. SJ 26322170 (quarry cliff base, Llanymynech), an impersistent thin MA.1.1 phase is developed intensely bioturbated

† Plate 14, fig. B

PLATE 14 Llanymynech Member.

A Cross-beds of shoaling 'type 1'. The junction of two opposed units is outlined. Note slight "convex-up" form of the cross-beds. Llanymynech Hill, G.R. SJ26822192.

B Minor 'sutured discontinuity surface' on top of a thin fenestrated calcisphere wackestone (MA.1). Old Quarry, Llanymynech Hill, G. R. SJ 26522183.

C Cross laminae of shoaling 'type 2' (oolitic). Llanymynech Hill, G. R. SJ 26762192.

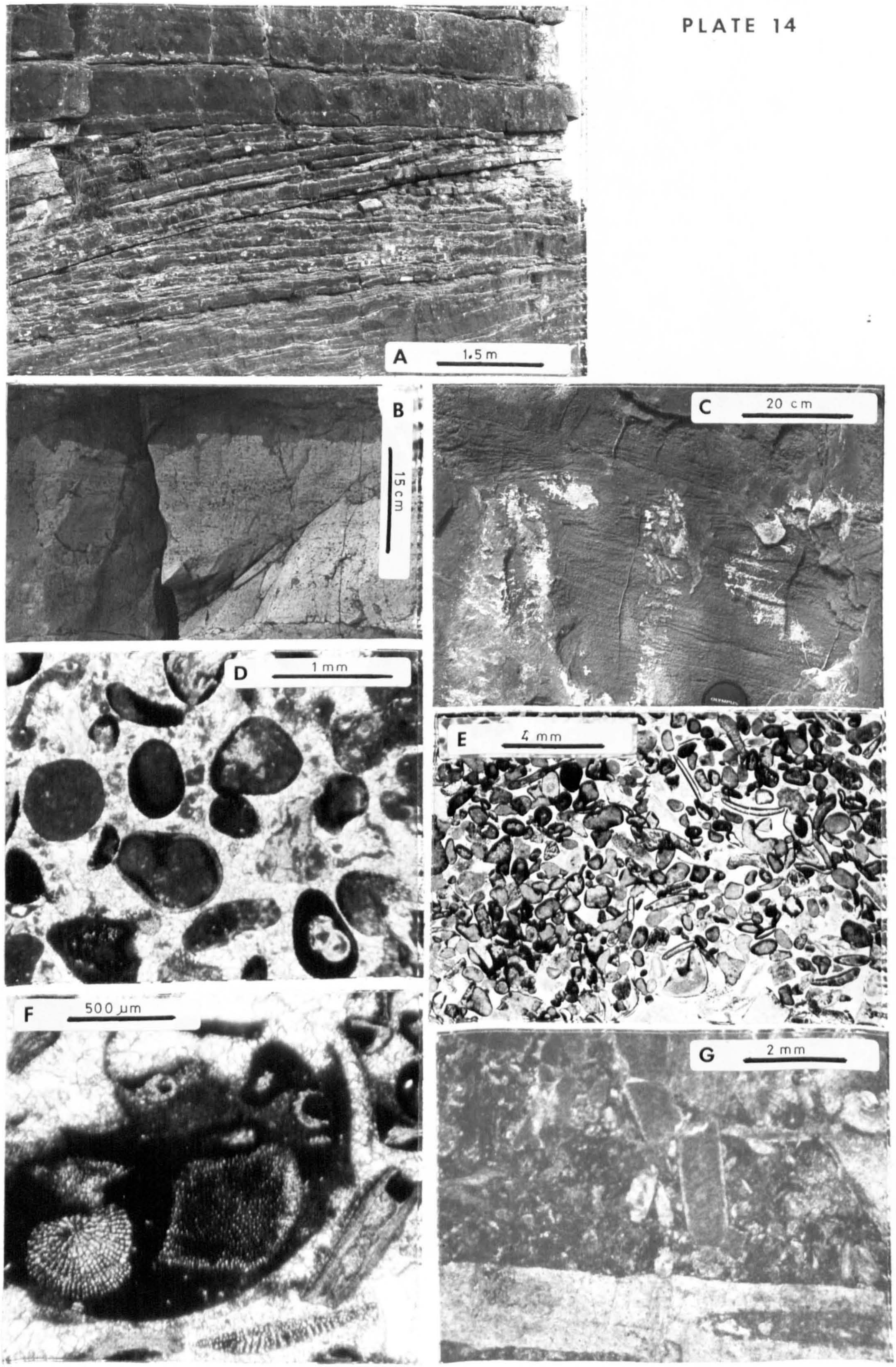
D Haematised superficial oolite. Note carbonate coating only partly replaced in grain at centre of photo , and irregularity in thickness of coatings. s/m 1149 G. R. SJ 26552184.

E Haematised superficial oolite, with mixed 'generations' of clasts. Note uncoated crinoid fragments in top right of photo s/m 1149.

G Bioclast packstone (MA.2.2) composed primarily of large fragments of echinoid spines. s/m 1146. G. R. SJ 26552184

F Haematised sediment internal to a gastropod fragment within the 'well washed' oolite of s/m 1149. Probably an intraclast (lithoskel).

PLATE 14



with Chondrites sp. (s/m 2570), infilled with the overlying MA.3 lithology.

No early cements nor peneocontemporaneous dolomites have been observed in Llanymynech Member MA.1 microfacies.

8.1.2. Alga packstone microfacies association (MA.2).

This microfacies association accounts for ~ 5% of the Llanymynech Member sediments, being more common in lower strata where there is a greater similarity to cycles of the Sychtyn Member. MA.2 microfacies occur as thin (5-20cm) laterally impersistent lensoid beds, often interlaminated and interbedded with finely bioclastic MA.3 sands. They comprise MA.2.1, MA.2.2. and MA.2.3 microfacies.

The calcareous shales are mostly maroon-grey in colour, and highly dolomitic, although they support a brachiopod-coral-sponge fauna.

MA.2.1 and MA.2.3 microfacies contain similar bioclast assemblages to Sychtyn Member MA.2 sediments and do not warrant further discussion here.

8.1.2.1. PETROGRAPHY OF MA.2.2. BIOCLASTIC PACKSTONE. This minor microfacies in the Llanymynech Member, associated with the other MA.2 microfacies, characteristically has a dominant molluscan (gastropod) component (e.g. s/m 1140), and has little recognisable algal microflora, although as with many examples of MA.2 lithologies in this member, micro-/pseudosparitisation has obscured the detail of the mud and silt grade matrix. As with MA.3.2 microfacies MA.2.2 is defined by having a > 20% recognisable, non algal bioclastic component. Echinoid spines are occasionally significant sediment contributors in this microfacies, (Plate. 14, fig. G).

Compaction features, both fragmented bioclasts and microstylolitic fabrics occur. No early diagenetic modifications have been observed.

8.1.3. Peloid Alga Grainstone Microfacies association (MA.3)

This microfacies association accounts for more than 90% of the Llanymynech Member sediments, occurring as cross bedded (angle of repose 5-15°; 'shoaling type 1') or massive internally cross laminated units, (shoaling 'type 2') traceable over good exposure along Asterley Rocks, Llanymynech Hill for about 1km as units rarely > 4m thick. Unlike any other Asbian cycles described herein, oolitic carbonates (MA.3.4) occur in these shoaling 'type 2' sediments (see section 8.4.1).

8.1.3.1. PETROGRAPHY OF MA.3.1 ; MA.3.2 and MA.3.3 These three microfacies are similar to those developed within the Sychtyn and Tynant deposits, varying from ~ 250μm sands to rudstones (MA.4.1). Two common compositional characteristics are:

1) Sediments of MA.3.2 (in 'type 2' shoaling units) are commonly dominated by fenestellid bryozoa, an exceedingly rare component of other Asbian sediments, (Appendix IV) up to 19% by volume e.g. s/m's 1167 ; 1170).
2) A local proliferation of echinoderm-rich sediment (also MA.2.2), including little-fragmented echinoid spines

Algae are correspondingly less significant in these sediments and beresellids commonly account for < 5% of the rock volume (cf. Tynant and Sychtyn sediments). As in Sychtyn Member MA.3 lithologies, Koninckopora occasionally accounts for about 15% of the sediment.

The sediments are commonly grey or rust -brown, the latter due to finely disseminated haematite. This is discussed in section 8.2. (e.g. s/m 1144)

8.1.3.2. PETROGRAPHY OF MA.3.4 OOLITIC GRAINSTONE The Llanymynech Member is the only Asbian unit recognised in this study with oolitic sediments well developed (about 5% of measured sections). These sediments occur within shoaling 'type 2' cross-laminated MA.3 units as the middle/upper phases of the Member cycles (see chart C and fig 19). The sediments are rust or grey-brown, either massive or stylolitically bedded, commonly with pervasive haematisation (section 8.2.1).

The laminae are oolitic/non-oolitic. Rudstone laminae (MA.4.3) are often associated.

The sediments are well-sorted, mud and silt deficient grainstones. Intraclasts peloids and bioclasts are well rounded, (200µm to 2mm sands. Oolitic coatings are superficial and range up to 150µm thick (Plate 14, fig. E). Coated grains range from less than 10% to greater than 50% of the sand fraction. Fenestellid bryozoa are common (e.g. s/m 1149), but bioclast compositions fall within the range of non-oolitic MA.3 microfacies in the Llanymynech Member.

Koninckopora is a significant algal sediment contributor, but is present only as micritised (and haematised) fragmentary thalli. The nature of the oolitic coating is distinctive. It is superficial and commonly highly ferroan, with disseminated haematite; forming concentric concentrations within the overgrowth, (Plate 14, fig D) or often ?replacing the overgrowth completely. Besides concentric tangential growth laminae, vague radial structure is visible (defining(?) secondary neomorphic fabrics). Carbonate overgrowth laminae do not show perfectly concentric growth, since in the initial growth stage, microscopic irregularities on the grain's surface are 'infilled'.

The overgrowth, therefore, thins towards grain surface 'highs'.

Occasionally two or more clasts are united by a single haematised overgrowth. Neither spastolithic (Hemingway, 1973) nor dripstone fabrics (Pullan 1967) have been observed.

It is not intended to delve into the problems of ooid formation here. Bathurst (1975, p.308) discusses many of these with respect to recent 'normal tangential' ooids, and highlights the mechanical/physiochemical or organic trapping models for their formation. Shearman et al. (1970) showed the importance of organic matrix in ooids in retaining the primary structure template during the calcitisation from precursor aragonite, although Sandberg (1975), suggested that many fossil ooids may have been primarily calcitic, invoking a Mesozoic-Cenozoic reduction in Ca/Mg ratio of the world's ocean chemistry.

Newell et al (1960) recognised that oolith growth commenced in waters $\geqslant$2m deep, reaching a maximum at 1.8m below low water in Bahamian environments.

8.1.4. Intraformational Conglomerate Lithofacies (L.7) and MA.4.

L.7 Lithofacies in the Llanymynech Member are limited to L.7.2 (matrix supported conglomerates). MA.4.1 are more common (bioclast peloid-rudstones). These two facies occur in similar settings as described in chapter 7 in the Sychtyn Member, with L.7.2 located as the basal lithofacies of cycles, with intraclasts of MA.1 lithologies derived from underlying cycles. At G.R. SJ 26542183 (Old Quarry, Llanymynech Hill - see chart C) L.7.2 with MA.1 clasts overlies a palaeokarstic surface, but there is no locally underlying MA.1, indicating that transportation over some distance (100's to 1000's of metres, possibly from contemporary Sychtyn Member sediments) occurred on the transgressive event.

Rudstones (MA.4.1) frequently occur associated with the oolitic cross-laminated and cross-bedded units, as thin layers, with only vague laminated internal structure, but whole-fossil (brachiopod) coquinas are rare.

8.2. EARLY DIAGENETIC FEATURES OF LLANYMYNECH MEMBER SEDIMENTS.

8.2.1. Ferruginisation.

Within all Llanymynech MA.3 microfacies haematisation has totally or partly affected varied proportions of the sand size grains. In fine grained MA.3.1 sediments, few grains are haematised, indicating the influences of transport-mixing (section 9. 3. 1), and supporting an early diagenetic origin of the haematite or haematite precursor. Haematisation is most widespread and intense within the coarse grainstones and oolites of shoaling 'type 2' sediments.

8.2.1.1 PETROGRAPHY Besides partly haematised oolitic coatings (section 8.1.3.2) bioclast microstructures are also replaced in a diffuse zone a few microns thick on clast surfaces, probably a consequence of selective replacement of an endolithic micrite rim. The inter-stereome honeycombs of echinoid spines are often partly haematised, as are micrite-filled bioclasts(?lithoskels), primary voids further indicating the replacive origin of the (now) iron-oxides. That some, but not all of the micrite-filled bioclast primary-voids and some but not all oolitic coatings are partly haematised infers that the replacement process was early and occurred on clasts prior to final deposition as a sediment of mixed 'maturity' (section 9.3. 1.2.) after much reworking. In s/m 1149, some intraclasts are noteably slightly haematitic. (Plate 14, fig. F) Neither chamosite, glauconite, nor siderite have been observed petrographically. Intergranular spar lacks the disseminated cryptocrystalline haematite.

8.2.1.2. SIGNIFICANCE OF FERRUGINISATION. Whether the original replacive iron mineral was haematite, requires consideration. Iron hydroxide (e.g. goethite) is a more probably primary mineralogy that is readily oxidised to haematite (Berner, 1969).

Recent ooids require high energy shoaling conditions for their formation, with sediment movement preventing them from adhering together, and allowing growth on all surfaces. The origins of oolitic iron formations are steeped in controversy. Sorby (1856) invoked a secondary, post-depositional ferruginisation, recently defended and amplified by Kimberley (1979), but a syndepositional origin has also been applied in many models, from early replacement to primary precipitation. Bradshaw et al (1980) in a reply to Kimberley (1979) recognised that more than one process may be responsible for the wide spectrum of oolitic iron formations in the geological record.

From the evidence above, a syndepositional replacement model is followed for these particular oolitic ironstones, although their occurrence in a sequence with varied emergence profiles does not completely preclude their formation by an illuviation-replacement process similar to the model erected by Kimberley (1979). The lack of a significant clastic component to the sequence alienates it from many clastic-carbonate oolitic ironstones discussed by Bradshaw et al (1980, p.295).

The "Bryozoa Bed" of the Lower Limestone Shales in South Wales is petrographically very similar to these North Wales oolitic and grainstone ironstones, with superficial haematised coatings, haematised primary bioclast voids and clear spar cement matrix. Greensmith et al (1971) proposed its syn-depositional formation. The Clinton Formation (Alling 1947) of the United States Silurian is also analogous in petrographic make-up. Adeleye (1973) described shallow marine goethite-haematite ooolitic facies from the Nigerian Cretaceous.

Thin interbeds of calcareous shales and shales in the Llanymynech Member are very reddened through disseminated haematite, presumably derived from weathering and transport of hinterland Lower Palaeozoic litharenites and slates (as readily observed in the Eglwyseg Formation). Ferruginisation is here considered as occurring due to:

1) Relatively greater movement of ooliths, and highly abraded grains leading to a greater time in contact with iron-charged marine waters, both within the surficial sediment laminae and as saltating bed-load. This process would not allow a grain to attain chemical equilibrium with its microenvironment. Each grain may have undergone many successive retransportation-redeposition processes before being finally buried;
2) Potentially iron-charged meteoric waters from the hinterland debouching into a partly enclosed marine environment (see fig 4) may have provided the necessary physicochemical parameters for replacement of carbonate by iron minerals, especially within ?algal/bacterial-infested oolitic coatings that may have induced reducing microenvironments.
3) Further support for this ferruginisation model is indicated by the ferruginous carbonates occurring predominantly in upper (shoaling 'type 2') phases of the Llanymynech Member cycles, which are interpreted as a regressive development (see section 8. 4.2.2) during which the body and depth of water within the Oswestry Embayment (see fig. 4) would have been substantially reduced. The ironstones are occasionally overlain by prominent 'form A' palaeokarsts. (Chapter 11).

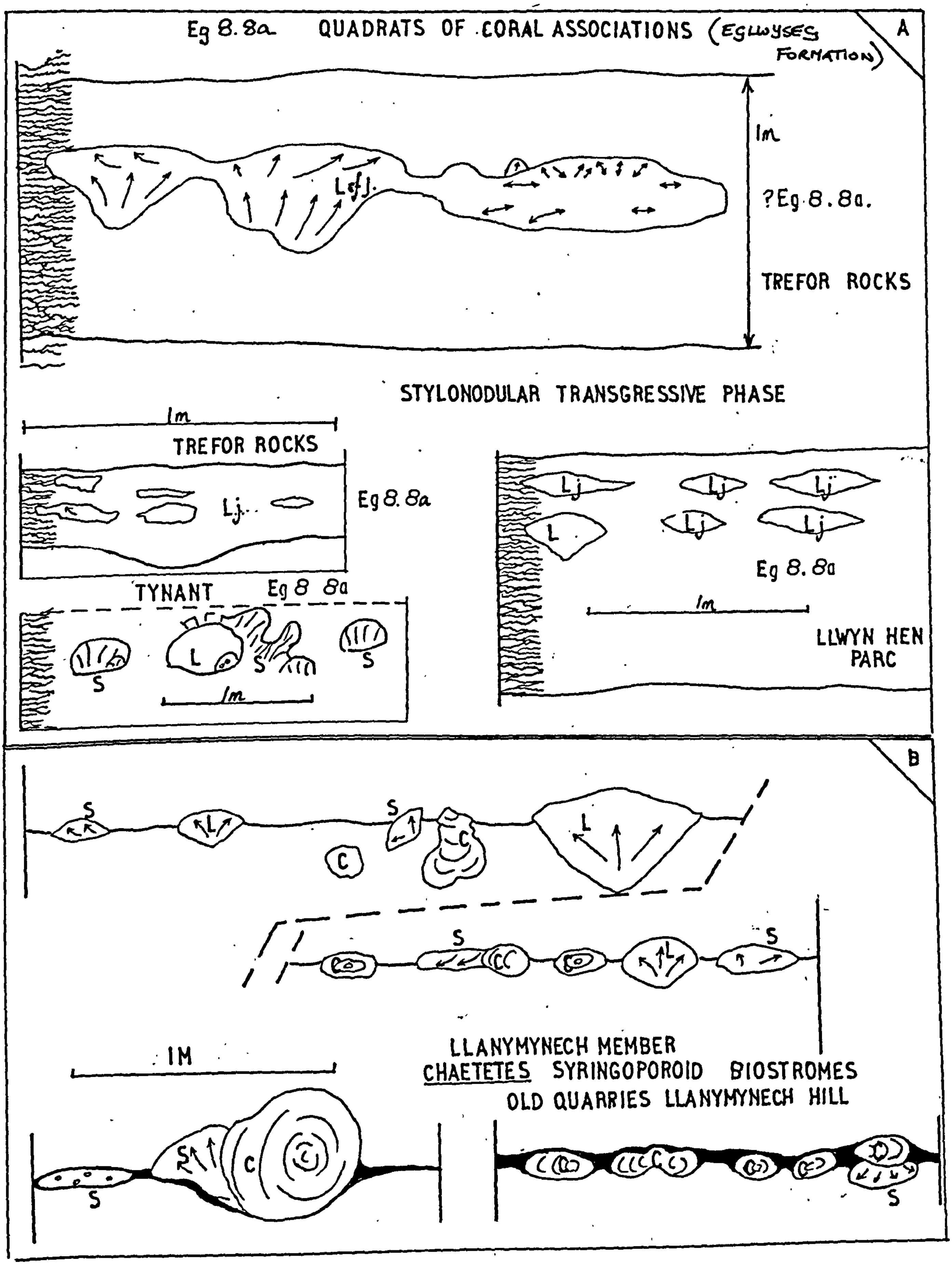
8.3. PALAEOECOLOGY OF LLANYMYNECH MEMBER SEDIMENTS.

8.3.1. Macrofauna Palaeoautecology.

8.3.1.1. PORIFERA Chaetetids are a dominant element of Llanymynech Member sediments. They occur in association with syringoporoids in laterally persistent biostromes (Fig. 55a) encrusting brachiopod valves, and in turn encrusted by syringoporoids. There is a particularly dense distribution of these biostromes towards the top of the "Old Quarry", Llanymynech Hill (G.R. SJ. 26522183). The colonies show successive overgrowth, with associated translocation of their growth centres (see fig 56) probably reflecting periodic disturbance of the colony by higher (than normal) energy bottom currents. A 'mushroom' growth, initiated on convex-up brachiopod valves (fig 56) is characteristic and may have been caused by one of two processes:

1) Current scour around the colony margins producing depressions into which the chaetetid preferentially grew (cf. flow principles of Richardson 1968) with eddy effects and faster flowing vortices allowing greater filtering capacities within the scour. Kershaw (1979) described similar phenomena associated with stromatoporoids from the Gotland Silurian.
2) Simple periodic overturing of the colonies, allowing an inverted mushroom growth.

The latter is most likely, but all the studied colonies exhibiting 'mushroom' growth were orientated convex-up (hydrodynamically most stable) and lack severe erosional breaks in growth. Growth occurred preferentially on the colony margins. The presence of few thin growth laminae orientated into the axial region of the mushroom (fig.56) indicates that growth occurred for a period of time into a central void, presumably inhibited through lack of nutrient-laden water flowage. This, therefore, imples that growth initiated upon convex-up brachiopods, with scour around colony margins enhancing marginal growth. Periodic overturning is then suggested by continuation of this marginal growth, and demise of the upper convex



S SYRINGOPORA
C CHAETETES
L CERIOID LITHOSTROTION
L.j. L. JUNCEUM

surface, Subsequent growth rendered the colony susceptible to further overturning (fig 56) . The axis, now infilled with sediment is a burial-diagenetic phenomenon. In the most extreme example of this growth upward bulging sediment laminae and microstylolites in this axial sediment body attest to this model. Syringoporoids preferentially grew on outer margins of the chaetetid 'mushrooms', that was probably an ideal substrate.

8.3.1.2. CNIDARIA. Solitary corals are very rare, represented by caninids and dibunophylloids, scattered throughout MA.3 microfacies, usually with eroded marginaria. The tabulate *Syringopora* is very common, occurring with *Chaetetes* colonies in laterally persistent biostromes. The colonies are mostly laid on their sides, indicating penecontemporary dislodgement, and occasionally encrust both chaetetid colonies and valves of productid brachiopods, although reciprocal encrustation by chaetetids is not known from this member (cf. Eglwyseg Formation). Colonies of *Syringopora* cf *reticulata* up to 50cm in longest axis, have a distinctive growth morphology, recognisable in colonies from the lower/ middle cycles of the Eglwyseg Formation also. It comprises one dominant growth axis, from which corallites diverge, producing a tall cylindrical colony that may have subsidiary axes off the main one. Storm disturbance may have readily broken off these subsidiary growth axes, thereby propagating new colonies, as reported by Perkins and Enos (1971) for broken and transported fragments of *Porites* colonies during large storms in Florida Bay.

BRACHIOPODA. The brachiopod assemblages are similar to those of the Sychtyn Member, with *Composita* sp rarely occurring in monospecific MA.4.1 coquinas. *Linoprotonia* and *Gigantoproductus* cf *maximus* are the two most common productids, both frequently inverted and with disaggregated valves.

GROWTH FORM OF CHAETETIDS FROM ONE BIOSTROME IN LLANYMYNECH MEMBER (SJ 26522182)

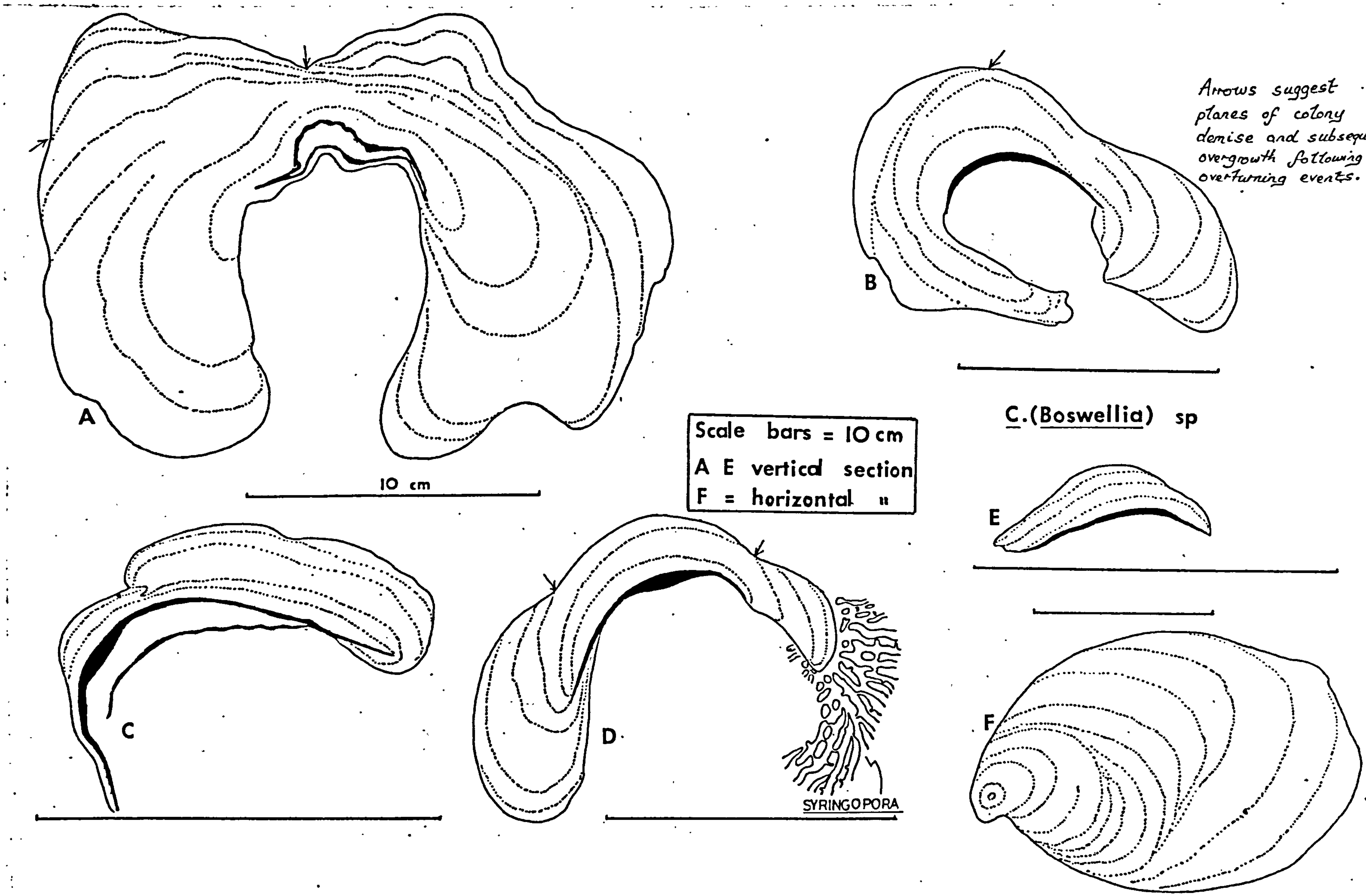


fig 56

8.3.1.3. <u>BRYOZOA</u> Both trepostome (e.g. s/m 1146) and more commonly cryptostome bryozoa, occur within MA.2.3 and MA.3 lithologies (e.g. s/m 1170). In particular 'type 2' shoaling sediments at times have bryozoans as the dominant clast type.

These bryozoan-bearing MA.3.2 and MA.3.4 microfacies within the shoaling 'type 2' sediments are unique within the Asbian sediments studied herein. Fenestrate bryozoa (<u>Fenestella</u> sp.) are the most common, occurring as colony fragments, up to cm-size, abraded and commonly associated with disseminated haematite and micrite rims, but locally accounting for over 30% of the sediment clasts by volume, and greater than half the bioclasts. In contrast, bryozoa in the overlying Brigantian Trefor Formation (section 10.1.3) are most abundant in MA.2 microfacies.

Mc Kinney and Gault (1980) studied the distribution of Chesterian (U.Dinantian) fenestrates from eastern United States, concluding that whilst some fenestrate forms preferred more protected bottom-waters, others thrived on shoals and shoal edges. As Mc. Kinney and Gault note (1980, p.128) zoarium morphology of the extant retiporiform bryozoa, a morphological analogue to Palaeozoic fenestrates, is well suited for high energy environments (Stach, 1936). Handford (1978) recognised echinoderm-bryozoan packstone/grainstone and bryozoan rich oolitic deposits from Upper Mississippian sediments (Monteagle Limestone), but noted their comparative absence in the interbedded wackestones and clay-shales. Stratton & Horowitz (1974) applied an experimental flume model to explain growth morphologies of flabellate fenestrate fronds, concluding that initially infundibuliform colonies grew to a flabellate form as a response to unidirectional currents, in a maximum current velocity of 1 to 3 cm/sec (very low).

Whether these Asbian fenestellids were infundibuliform or flabelliform is unclear, but if their tolerance was limited to 1 to 3cm/sec. currents, then it implies that higher velocity fragmenting and disrupting currents were exceptions and not the norm. As bottom currents of $\geqslant$ 35cm/sec are required to move 750µm sand grains (Hardie and Ginsburg, 1977) and this was achieved frequently in MA.3 microfacies, a velocity approaching 3cm/sec is an unrealistic norm. The lack of prolific whole and slightly abraded bryozoans within associated MA.2 sediments further indicates that the bryozoans were indigenous to MA.3 substrates, and not transported far (assuming lateral equivalence of MA.2 and MA.3). In conclusion, therefore, some Asbian fenestrates preferred to dwell adjacent to and upon 'sorted' sand substrates, but periodic turbulence and movement of bottom sediments fragmented and abraded these colonies without significant transport.

8.3.2. Petrographic Diversities.

MA.3 sediments of the Llanymynech Member have similar standardised petrographic diversities to MA.3 sediments within the Sychtyn Member, with total diversities averaging 19 ± 3 (see table 5). Algal diversities average 7 ± 2 and microfaunal diversities 6 ±1. Macrofaunal diversities are higher than the Sychtyn Member due to the presence of bryozoa, and a locally prolific mollusc assemblage. This may reflect a greater sediment mixing and more proximal bioclast production (and fossil growth) than in the more shoreward Sychtyn Member depositional belt (section 2.4.4.).

8.4. SEDIMENTARY MODELS FOR THE SHOALING FACIES.

It is pertinent to discuss the sediment geometry and member relationships here as these have great bearing upon the palaeoenvironment interpretation of the Llanymynech Member.

Fig. 4, indicates the palaeogeographic setting of the

Oswestry area, showing that the Llanymynech Member occupies a 2km tract at the southern termination of the Whitehaven Formation outcrop. The shoaling nature of much of the Llanymynech Member in comparison to the laterally equivalent (and more landward) Sychtyn Member is indicated above and shown in Chart C. Whether this exhumed outcrop pattern reflects an original geometry of the shoaling area, remains unclarified. If it is, then the area subparallels the projected coastline, and lies 'seaward' of the embayed Sychtyn Member depositional area. There is no borehole nor geophysical evidence to show whether any palaeoslope existed seaward of the shoaling facies (towards the Craven Basin), that may be detected as a transition to sediments of more basinal facies.

Comparison of the cycle form between the Sychtyn and Llanymynech Members (figs.28 & 29) show that whilst MA.1 sediments are being deposited in the former, MA.3 shoaling facies are their likely Llanymynech Member chronostratigraphic counterparts. Many comparable situations, both ancient and modern have been documented.

The shoaling facies are of two distinct types related to cycle form:

Type 1. A lower cross bedded finely peloidal-bioclastic MA.3.1 dominated facies.

Type 2. An upper oolitic and coarse sand MA.3.2 and MA.3.4 dominated facies, with internal cross lamination.

8.4.1. <u>Internal form and Geometry</u>.

8.4.1.1. <u>SHOALING TYPE 1 UNITS</u>. The cross bed sets are up to 3m thick, occurring towards the base of minor cycles. They vary from a repose angle of 2 or 3° to 15°. Individual units are traceable along Asterley Rocks for up to 1km. Foresets are rarely convex-up (Plate 14, fig.A). Beds are 5 to 50cm thick, mostly of fine parallel laminated

MA.3.1 and MA.3.2 sands with argillaceous interbeds. Neither topsets nor bottom sets are apparent, and bounding surfaces are sharp A northerly to easterly foreset trend appears to dominate. No early cement features have been observed, nor is there evidence of any sediment trapping organisms (forming 'bafflestones').

At G.R. SJ 26762191 (Llanymynech Hill) a ?channel structure is visible in section, cutting into underlying cross bedded finely bioclastic MA.3.1 (fig 57).

8.4.1.2. SHOALING 'TYPE 2' UNITS.† These more massive, internally cross-laminated sets up to 4m thick, show evidence of migrating bidirectional ?megaripples in section (fig. 57). Similar features occur in upper cycles of the Sychtyn Member (Craig y Rhiw, G.R. SJ 23602595, capped by a thin MA.1 phase underlying a palaeokarstic cycle boundary).

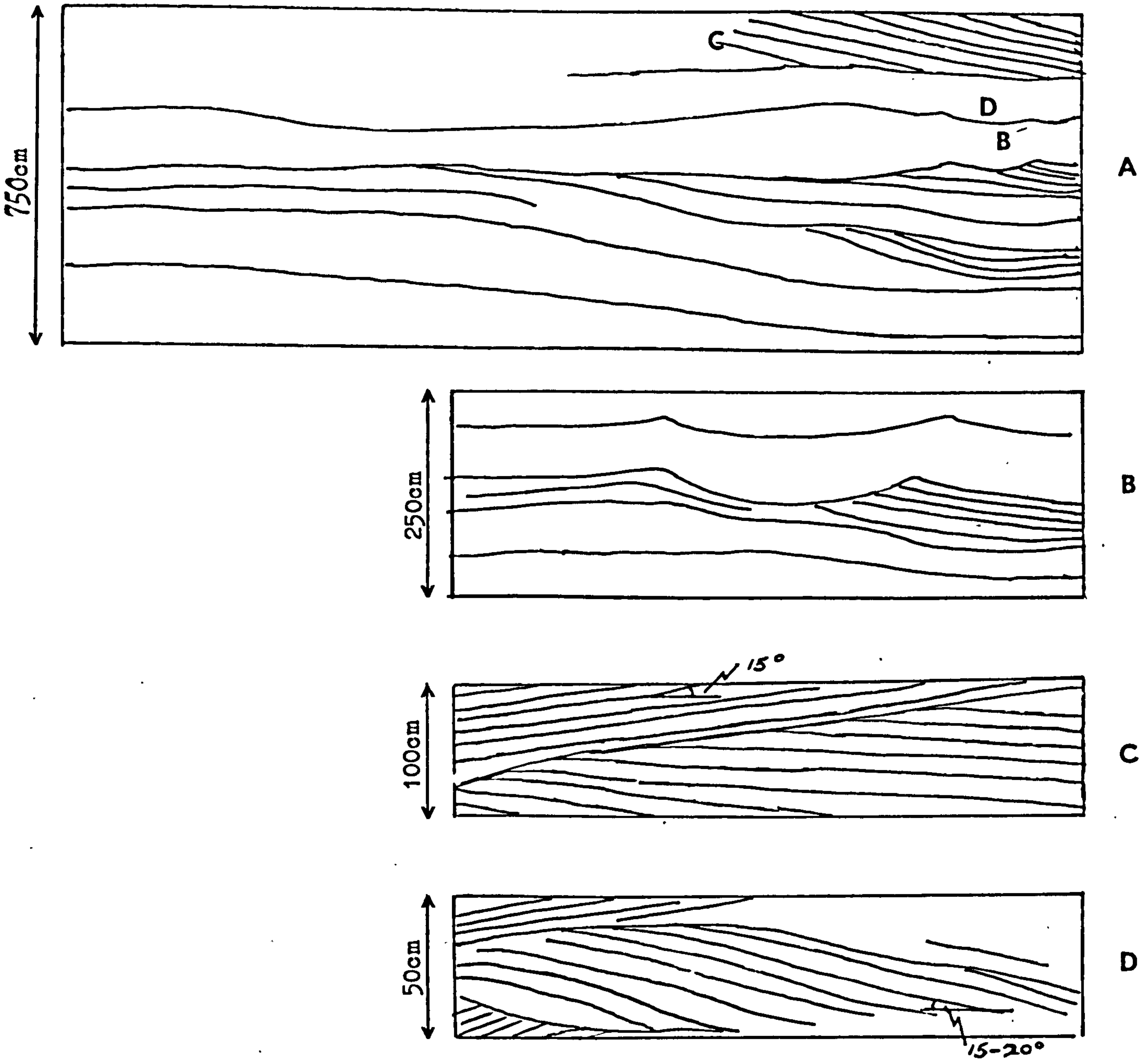
They are dominantly intensely haematised, well washed and rounded coarse MA.3.2 and MA.3.4 oolitic sands, with high crinoid and bryozoan content, but significantly lacking calcispheres and beresellid algae. These units occur towards minor cycle tops representing regressive phase sedimentation.

8.4.2. Recent Analogues..

8.4.2.1. TYPE 1 ANALOGUES Recent cross bedded carbonate units, both with and without associated oolites are formed in shoaling shelf environments, typically as elongate sandbodies. Ball (1967) classified such Bahamian sand bodies as marine sand belts, belts of tidal bars and eolian ridges.

Oolitic marine sand belts occur near slope breaks, with sand movement across this belt, forming spillover lobes. The Bahamian examples have rippled, locally tidal-emergent upper surfaces with a relief of up to 12 feet. Sediments move by both tidal and wind-

† Plate14, fig. D.



SKETCHES OF BED MORPHOLOGIES AND INTERNAL STRUCTURES OF MA.3 SEDIMENTS - LLANYMYNECH MEMBER (G.R.SJ.26762191).

A) Cliff-face bed structre. Note presence of irregular lensoid beds and poorly defined nature of some ?cross-beds, probably a 3D effect.

B) Enlargement of bed structure shown in A. Note possible bedform or channel in section.

C) Foreset pattern in large-scale cross beds.

D) Internal cross-lamination, ?section of migrating megaripple.

Fig 57

driven currents and support a molluscan dominated macrofauna. Ball (op. cit. p.107) recognised a two-fold structure to these bodies: a basal set of large or medium scale convex-up cross-beds overlain by small scale cross-bed sets, representing initial migration of the sand belt and subsequent reqorking on their tops, with migration perpendicular to belt trend. The ooliths are mostly well developed and are most concentrated in the basal larger scale cross-bed sets.

Tidal bar belts and Eolian ridges are less comparable, due to their high density of cross-belt channels and large scale cross-sets respectively (Ball op. cit. p.129), although minor tidal influences are plausible, from the limited palaeocurrent data.

'Bank units' associated with Shark Bay subtidal sediments (Hagan and Logan, 1974) also develop features similar to the cross-bedded units of the Llanymynech Member. They have acretionary slopes from 2° or 3° to 15°, comprising of massive or thin bedded wackestones packstones and grainstones. Fringing, barrier and patch bank units have been recognised. They form through trapping of sediments by seagrasses and are modified by the influence of waves, tidal currents, and substrate gradients.

8.4.2.2. TYPE 2 ANALOGUES. Recent platform interior sand belts are worthy of close comparison with 'type 2' facies. Ball (1967) exemplified the large expanse of sands west of Andros Island, occupying a location intermediate between the shelf slope and low energy inshore sediments. Periods of emergence are fossilised within the sequence, and coarse lags attest to periodic storms, but for the most part the sands contain a significant mud (and silt) fraction. Superficial oolitic coatings are common. These sediments correspond to the unstable sand habitat of Bathurst (1975).

The formation and distribution of oolitic coatings on Bahamian sediments reaches a maximum 1.8m below low water (Newell et al 1960), and falls substantially towards deeper waters, where they concluded (op. cit) that ooid growth commenced. The early acicular aragonite fringes common on Bahamian shoals (Davis 1979) have no fossilised counterpart in the Llanymynech Member.

8.4.3. Fossil Counterparts

Wilson (1975, p.283) described shoaling oolite-grainstone cycles from the Mississippian of the Williston Basin, Montana, comprising alternating dark argillaceous limestone with oolites. These cycles bear striking similarity to those of the Eglwyseg Formation, with the prominent development of an oolitic/grainstone regressive upper phase.

Handford (1978) recognised an oolitic tidal-bar complex in the Mississippian Monteagle Limestone (Alabama), his model based on evidence of a bipolarity of palaeocurrents in the oolitic facies, and a linear bar-like 3D geometry of these bodies, trending along the strike of the complex.

8.4.4. Interpretation of Llanymynech Member Shoaling Facies.

The model adopted for the sand bodies described here cannot be a direct analogy of these recent and fossil counterparts, although specific points of each model fit. Table 6 summarises the interpretation to the data presented here.

8.4.4.1. SHOALING TYPE 1 FACIES. A 'bank' formation similar to that of Shark Bay is not readily conceivable for the Llanymynech sediments, through ?lack of initial surface relief, and requirement of a sediment-baffling organism (not bryozoan in shoaling type 1). A marine sand-belt interpretation is preferred here. Table 6 summarises the most probable interpretation.

The lack of oolites within the shoals, and association with the transgressive event of the minor cycles suggests they formed in water depths > 2m. They formed in a relatively low energy environment, most probably in response to fine sand transport from inshore (Sychtyn Member type) environments.

The fine sand nature, and occasional convex-up form of the cross-beds suggests waning current strength and reduced sediment supply modified their form.

8.4.4.2. SHOALING TYPE 2 FACIES. These cross-laminated units of Llanymynech Member cycles often intensely haematised, reflect small scale migratory bedforms developed above 'type 1' shoals during regressive phases. Type 2 facies are not formed by reworking of type 1, shown by differences in grain size and sediment composition. Occasionally emergent (producing palaeokarsts and calcrete phenomena as evidence), these sediments were submarine rather than either beach ridge or aeolian, indicated by the presence of little disturbed sponge-coral-brachiopod assemblages.

They probably formed in waters ≤ 2m deep (from oolite presence), in part affected by ?tidal currents. They are in part laterally equivalent (?) to the regressive peritidal suites of the Sychtyn Member (see table 6).

It must be emphasised that these shoaling facies are interbedded with MA.3 sediments typical of Sychtyn Member minor cycles (chapter 7), that show no migrating shoal characteristics and that rarely are both shoal forms present within the same minor cycle. (Chart C)

TABLE 6

SUMMARY INTERPRETATION OF LLANYMYNECH MEMBER SHOALING FACIES

	FACTS	INTERPRETATION
Lower cross-bedded phase. (Type 1)		
1	Large scale foresets; assocition with indigenous benthos; limited (?) lateral extent, seaward of sand-mud cycles	Either : marine sand belt; tidal bar sand belt; or patch / barrier'bank'complex (closest modern analogues).
2	2m thick foreset units have sharp bounding surfaces, are only locally oolitic, dominated by finely bioclastic peloidal grainstones, with occasional rudstone laminae. Lack of cross-lamination on any scale. Varied palaeocurrent directions, ?mostly towards the N and E. Argillaceous interbeds. Bedding between 5 and 30cm thick.	Primary relief of $\geq$ 2m, commencing as a non-gradational event. Non-oolitic nature suggests water depth $\geq$ 2m. Stenohaline moderate to low energy environment for the most part. Periods of sand movement (mostly sorted fine sand) parallel with and seawards of depositional strike. Varied foreset directions imply a 'lobe stratigraphy'. Varied foreset directions are only evidence(?) of any tidal influence. No evidence of significantly large and common channels. Upper surface apparently not emergent. Terrigenous clastic input periodically dominates carbonate sedimentation. No evidence of initiation through primary relief.
Upper oolitic phase. (Type 2)		
1	Commonly oolitic (superficial),'well washed', abraded and coarse sands. Often cross laminated with 50cm relief foresets. Indigenous benthos.	Deposition in shallow waters($\leq$ 2m), with nucleation of ooids. Stenohaline, moderate to high energy for the most part. ? Blanket shelf sands.
	Palaeocurrent directions dominantly northeastwards.	Megaripples moved NE across the sand belt, and (?) along its axis. Locally bidirectional megaripples suggest tidal influences.
	Rudstone laminae throughout.	Rudstones suggest storm events.
	Units rarely >2m thick, commonly <1m.	
2	Contemporary'haematisation', and admixture of different generation clasts.	Ferruginisation of some sand (not necessarily haematite originally - iron hydroxides/ silicates and even glauconite are possible primary mineralogies). Reworking of sands (?mostly local) leads to more 'mature' composition ranges.
3	Overlain by palaeokarst and rare calcrete phenomena	Emergence on maximum regression, exposing sands to karstic dissolution, presumably following and / or contemporary with partial lithification (of which there is now no evidence).

9. MICROFACIES AND PALAEOENVIRONMENT OF THE EGLWYSEG AND LLYNCLYS FORMATIONS.

9.1. MICROFACIES ANALYSIS

9.1.1. Microfacies characters of transgressive suite sediments.

The basal phases of minor cycles in both the Eglwyseg and Llynclys Formations comprise a sequence of argillaceous and thin bedded or stylonodular packstones and grainstones, containing a diverse stenohaline fauna. Sediments of MA.1 microfacies association are rare, restricted to lower cycles of both the Eglwyseg and Llynclys Formations.

9.1.1.1. MA.1 (CALCISPHERE WACKESTONE) SEDIMENTS. MA.1 units occur as thin ($\leqslant$ 1m) beds within 'Eg.1.2', 'Eg.3.4', 'Ly. 1.2' and 'Ly.2.3'. They have similar petrographic characteristics to the MA.1 sediments described in section 6.2., dominated by MA.1.1 (calcisphere mudstone to packstone). MA.1.4 (laminated beresellid-calcisphere mudstone to packstone) has been recognised at Nant Mawr Quarry (s/m 1109, Llynclys Formation).

These minor units of MA.1 sediments have emergence features associated with their upper surface (S. D. surfaces (section 11.1) and tubular calcrete structures) indicating that although they were deposited within a packet of transgressive sediments periodic regression and emergence ensued, in a similar manner to the underlying Tynant and Whitehaven Formations.

9.1.1.2. MA.2 (ARGILLACEOUS ALGA PACKSTONE) SEDIMENTS. ((Fossiliferous shales (L. 62) occur as impersistent, poorly exposed units <2m thick. The basal phase of cycle 'Eg. 2. 3' at Bron Heulog is the thickest development exposed within the Eglwyseg Formation. As homogeneous grey shales, grading through calcareous shales into argillaceous packstones they contain a brachiopod dominated assemblage (primarily Linoprotonia, Rugosochonetes, spiriferids and Composita), with little or no evidence of reworking or fragmentation of the valves. Gigantoproductus cf maximus occurs infrequently, being more common in higher strata.)).

The calcareous shales (MA.2.3), however, contain a more diverse coral brachiopod fauna with fasciculate lithostrotiontids and syringoporoids commonly flat and crushed along bedding planes (e.g. cycle 'Eg. 3.4.', Trefor Rocks, cycle 'Eg.4.5' Tynant, cycle 'Ly.2.3', Nant Mawr and Whitehaven) along with crinoidal debris in various stages of disaggregation.

Beresellid packstones to wackestones (MA.2.1) are common thin-bedded basal phases, similar both in external and petrographic appearance to the Tynant Formation MA.2.1 (section 7. 2. 1). They have identical compositions to Tynant Formation sediments, occurring throughout the Eglwyseg Formation, with the dominant beresellid Kamaenella sp. (see Appendix IV e.g. s/m 044 and 043). Macrofossil bioclasts are subordinate, and in s/m 053 (Eg.10b / Tfl, Tynant) detrital quartz (25 to 75µm) accounts for 25% by volume of sediment. These sediments plot towards the A end member of the C-A-M ternary diagram (section 2. 1.2).

Bioclastic packstones (MA.2.2) are minor microfacies of the basal thin bedded MA.2.1/ MA.3.1 dominated transgressive phases. They are characterised by > 20% by volume macrofossil bioclast contribution (e.g. s/m 306, 'Eg.8. 8a'). Neomorphic fabrics have commonly obliterated much of the fine detail of these sediments (section 9.1.2.)

9.1.1.3. MA.3 (ALGA PELOID GRAINSTONE) SEDIMENTS. In basal phases of many Eglwyseg Formation cycles MA.3 sediments are the dominant microfacies association, occurring either: 1) in 5 to 50cm beds of alga peloid grainstones (MA.3.1) or: 2) as interbeds and laminae within the argillaceous stylonodular basal phases:

1) These MA.3 sediments are similar both petrographically and in outcrop to the well bedded MA.3 sediments of the Tynant Formation, lacking the pale massive character of MA.3 dominated upper phases of Eglwyseg and Llynolys Formation cycles. They are medium to pale grey with either smooth flat bedding planes, defined by thin argillaceous partings, or stylolites. Their darker colouration compared to overlying massive bedded phases is probably a product of slightly greater insoluble clay residues and / or lack of fine crystal silt 'cement' (see on). Konincopora and beresellid algae are ubiquitous in these sediments throughout the Eglwyseg Formation but they retain a similar uniform petrographic fabric to the MA.3.1 sediments of the underlying Tynant Formation.

2) The neomorphic microspar/pseudospar (sensu Bathurst, 1975,p.566-7) fabric of many stylonodular basal phases obliterates primary fabrics, many of which may have been MA.3.1 sediments, although commonly the relict clasts have a grain float texture indicative of original wackestone (MA.2.1 or MA.2.2) texture.

9.1.1.4 L.7 (INTRAFORMATIONAL CONGLOMERATE) SEDIMENTS. Clast supported intraformational conglomerates occur in basal transgressive phases as palaeokarst relief filling sediment. The classic locality for this is G. R. SJ24004277 Bron Heulog, above 'Eg.6' palaeokarst. The sediments are unbedded and polymict, but clasts up to 20cm diameter are stained ochres and reds, with haematite impregnation. At this particular locality, the sediments grade upwards into L.6.1 (unfossiliferous maroon mudstone) with conglomeratic layers that contain very large haematised ooids ($\leq$ 3mm diameter and $\leq$ 750µm oolitic coating (Plate 15 fig. A)), fragmentary bioclastic material and intraclasts of MA.3 lithologies up to 6cm diameter.

Thin laterally impersistent bioclastic rudstones (MA.4.1) occur within both bedded MA.3.1 and stylonodular phases.

9.1.2. Neomorphic Fabrics of Stylonodular Sediments.

Stylonodular sediments as cycle basal phases are either intensely neomorphosed, containing vague relicts of original sediment texture, or have a uniform neomorphic texture, with 20 - 70 µm crystals defined by undulose surfaces with thin dark coatings, probably relicts of insoluble residues. This latter fabric is most common in thin (5-15cm) parallel bedded sediments, interbedded with clays, and grading vertically and laterally into more truly stylonodular sediments. These well developed interbeds may be either maroon, grey or ochre clays, but are mostly less than 2cm thick.

Below are listed the characters of aggrading neomorphic textures apparent in these stylonodular phases:

1) A pervasive microsparitic neomorphism of mud- and silt grade matrix and cement, producing a pale grey to yellow coloured equigranular mozaic. This neomorphism may initiate at points throughout the sediment, producing coalescing mozaic fronts.† Each 'patch' of neomorphism varies from a few millimetres to 3 or 4cm diameter. The neomorphism affects different clast types to varying degrees. Brachiopod fragments appear most resilient to the process, mostly remaining unaltered, whilst other clasts are retained only as vague ghosts (defined by inclusions and crystal size) in the neomorphic zone.
2) As the neomorphic fronts coalesce, the growing crystals commence to increase to a millimetre-size producing a radial fibrous

† Plate 15, fig. C

PLATE 15 Eglwyseg Formation.

A Large, haematised oolith retaining vague concentric structure. Infill of palaeokarst relief of Eg. 6., Bron Heulog. G.R. SJ 24024275.

B Crystal silt within foraminiferal test. Note adjacent tubular-septate algae (A) with syntaxial cement mm s/m 039, Tynant, G. R. SJ 22104526.

C Coalescing neomorphic 'nodules' within stylonodular phase. Dark internodule streaks are clay-rich. S/m 316, Stylonodular phase of Eg. 5. 6., Trefor Rocks. G. R. SJ 22644360

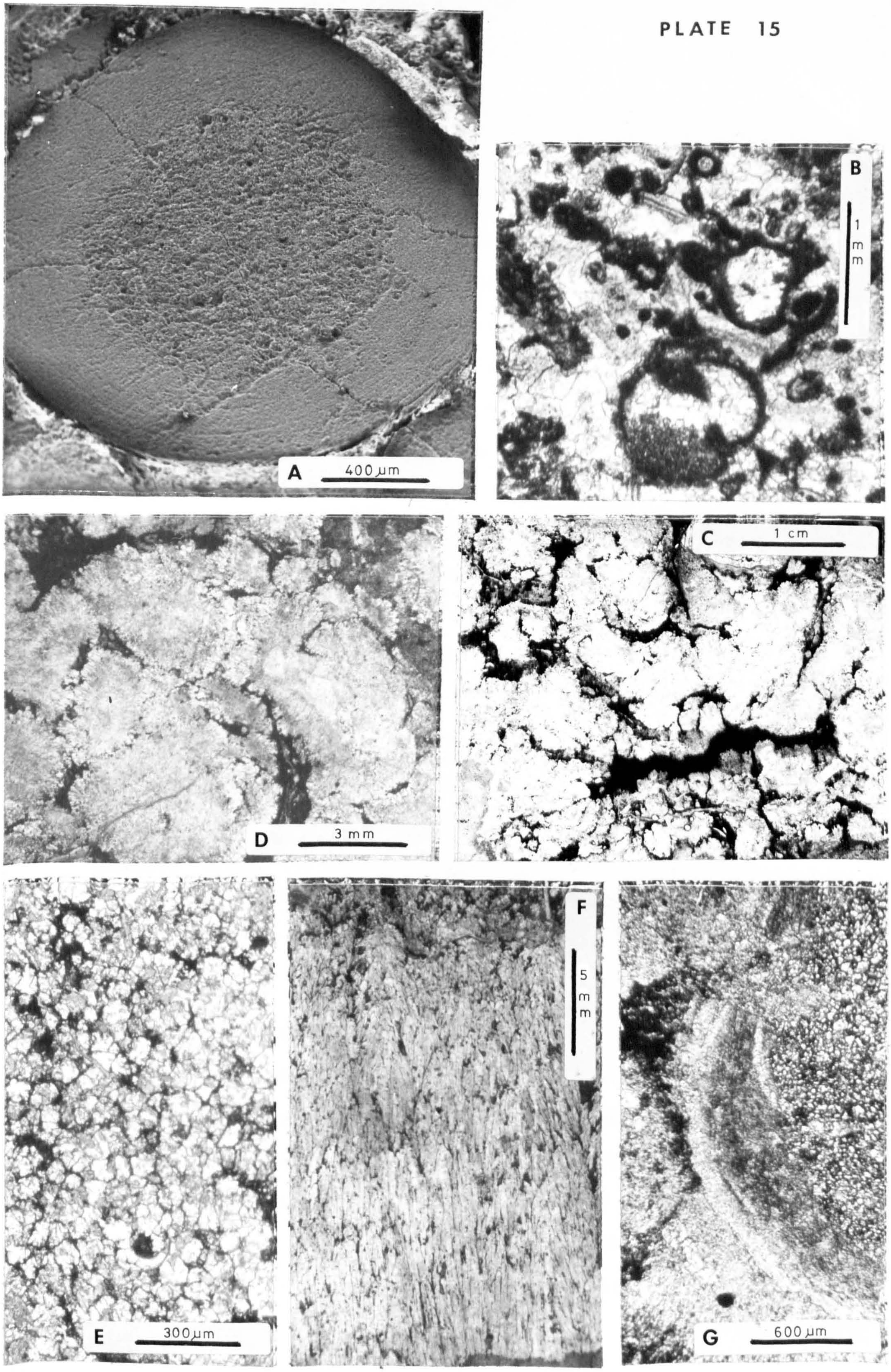
D Close up of fig. C. Note coarser neomorphic fibrous fabrics on 'nodule' margins.

E Pseudospar neomorphic texture of stylonodular sediments. Note vague relicts of ?calcisphere at bottom centre.

F Neomorphic fibrous calcite 'palisade'. A late neomorphic vadose effect? s/m 316.

G Neomorphic micro/pseudosparitisation around a syringoporoid corallite. Note coarser neomorphic fabric within the corallite s/m 316a.

PLATE 15



texture, resembling poorly developed cone-in-cone. Residual whisps of insoluble residue rich sediment, with dolomite crystals (20 - 100μm) remain. Clasts in these residual whisps show a flowage fabric around the growing microspar/pseudospar 'nodules', indicating that neomorphic growth was partly a selective recrystallisation process that 'pushed apart' the non calcitic components, rather than incorporate them as inclusions within the aggrading mass. The coarser neomorphic fabrics tend to occur in embayed margins between coalescing nodules and within intraparticulate sediment/void. (Plate 15 fig D &G). This may be due to a pressure shadow effect in these areas enhancing the growth process. In sediments with grain support texture (of larger (brachiopod) bioclasts) the neomorphic fabric may be coarse† ($\geq$ 50μm), possibly reflecting pressure shadow effects produced by these clasts acting on the matrix.

Similar neomorphic textures (pseudosparitic, and commonly of ferroan calcite) occur marginal to, and along the length of, the stylolites defining the stylonodular character, and also contain dolomite within the thin insoluble layer associated with the stylolite.

<u>INTERPRETATION.</u> Longman (1977) proposed that microsparitisation was induced by loss of Mg ions from calcite micrite, due to sump effects of surrounding clays, especially montmorillonite and chlorite, attracting the Mg ions. With reduced Mg lattice impurities, the remaining calcite is more susceptible to dissolution (and reprecipitation in a thermodynamically more stable structure and orientation). It is significant to note that in these Asbian sediments argillaceous whisps, often as stylolite flasers, and

† Plate 15, fig. E

insoluble rich fringes to neomorphic zones, commonly contain dolomite. Illite, the most common clay mineral of these sediments does not have the Mg++ absorption capacities of the smectites, but its presence may induce adsorption of Mg++, from which dolomite may more readily precipitate.

Although stylonodule formation is enchanced by the presence of mottle 'pseudobreccia' (section 9. 2. 2) there is no evidence to support their formation due to subtidal lithification at an early diagenetic stage, as recognised by Jones et al (1979) in the Upper Silurian of Arctic Canada. They (op. cit.) cited evidence of burrowing both within and avoiding the nodules as evidence of their early formation.

Nichols (1966) described the petrography of nodular limestones from the North Wales "D_2" (Brigantian) exhibiting similar neomorphic textures to those described here, although he did not recognise a neomorphic origin of the 'calcite silt' (microspar).

9.1.3. Late Stage Neomorphism - Palisade Growth.

At the base of many thin stylonodular 'beds' in more argillaceous sediments (e.g. above 'Eg.5' Trefor Rocks, 'Eg.6' and 'Eg. 7', Bron Heulog) is a downward projecting growth of fibrous palisade calcite (Plate 15 fig.F). Crystal growth initiated from an irregular surface within pseudosparitised sediment, but growth was in a downward direction only, with successive crystal generations increasing in size to a maximum of about 1cm long. These masses may reach 2cm in thickness (e.g. s/m 316, above 'Eg.5' Trefor Rocks.), are dark and lack fine inclusion-rich laminae within. Bladed terminations 'grow' into finer neomorphic fabrics.

INTERPRETATION. These growths superficially resemble speleothemic calcite. Kendall and Broughton (1978) described inclusion-laminated palisade calcite speleothems. They suggested that crystal growth and incorporation of planar inclusion laminae are the result of coalescing small syntaxial crystallites on an essentially planar growth surface. The gravity controlled morphology of these palisades implies deposition from, or under the influence of vadose waters. The growths are not transected by either neomorphic fabrics, stylolites or late stage veins. This indicates that they are late stage diagenetic features, and are probably associated with post Palaeozoic uplift. It is unclear whether the growing margin replaces or displaces underlying carbonate. This fabric superficially resembles cone-in-cone structure.

9.1.4. Microfacies characters of regressive suite sediments.

Regressive suite sediments form upper phases of the minor cycles, dominated by massive bedded MA.3 lithologies, and include much pseudobreccia mottling. Bedding is vague to absent, defined by either prominent denticular stylolites or petrographically distinct layering, visible in the field by layers of different shades of pale grey and brown. It is the lack of bedding planes and generally paler colouration that distinguishes this from underlying MA.3 sediments associated with the transgressive suite.

9.1.4.1. MA.3 (ALGA PELOID GRAINSTONE) SEDIMENTS The sediments are mostly homogeneous finely bioclastic sands, but centimetre laminated parallel and cross laminated units do occur. Darker layers represent 'cleaner washed' grainstone layers.

Beresellid algae, including such forms as Kamaena and the ubiquitous Kamaenella , at times contributing over 50% by volume of the sediment, but decreases in coarser, well sorted grainstones. Koninckopora is also an ubiquitous alga, varying from being a significant individual contributor (10% by volume) to 1%., and from intensely micritised fragments to large fragments of little abraded thalli.

Endothyrid, textularid, and archaediscoid foraminifera are other major microfossil sediment contributors.

MA.3.1 sediments are the most prolific, with clast sand size range between < 250µm to > 1mm. Varying proportions of coarser fragmentary bioclastic material floats within this matrix. Much of the 'cement' is silt grade calcite (non ferroan) and syntaxial spar, with some spar druse that is commonly zoned to centres of more ferroan calcite (Potassium ferricyanide/Alizarin Red-S stain.) .

MA.3.1. sediments are often interlaminated with MA.3.2. (Bioclastic Grainstones), especially lenses of crinoidal grainstone, 1 to 4cm thick.

Fine (400µm) well sorted peloidal grainstones (MA.3.3) that have a 'porcellanous' character are rare microfacies within the massive bedded units (e.g. s/m 1246, Plate 16, fig. A). The peloids are mainly irregular partly micritised bioclast fragments. No pellet (faecal) sands have been found. In s/m 1246 burrow structures (5mm diameter, subvertical) are filled with more silt-grade material, producing a fabric relationship of opposite form to that of pseudobrecciated sediments (see on).

9.1.4.2. MA.4.1(BIOCLASTIC AND PELOIDAL RUDSTONE) SEDIMENTS. Brachiopodal rudstones and floatstones (the latter with MA.3 matrix) occur commonly within the Minera sequence (see Chart E) between 'Eg.1' and?'Eg. 6'., and rarely within the massive MA. 3.1 dominated phases as thin, laterally impersistent lenses up to 30cm thick; The Minera examples comprise convex-up single-valved Linoprotonia, with much primary void/umbrella cavity†, but also have a diverse assemblage of bioclastic debris in varied states of abrasion and fragmentation, including fasciculate lithostrotiontids and chaetetid colonies. Many of these clasts have thick ?exolithic algal micrite rims, and are associated with minor proportions of Lower Palaeozoic extraclasts, and intraclasts up to 3cm diameter.

9.1.4.3. MA.1 (CALCISPHERE WACKESTONE) SEDIMENTS. The upper phase of 'Ly. 1. 2', (Chart C) is the only recognised MA.1 capping to this style of minor cycle, itself terminated by a prominent palaeokarst, and reflecting the progressive change from underlying Sychtyn Member to Llynclys Formation cycle form,

9.2. EARLY DIAGENETIC PHENOMENA OF EGLWYSEG AND LLYNCLYS FORMATION SEDIMENTS.

9.2.1. Cement Fabrics.

9.2.1.1. ACICULAR CEMENT RIMS. These are rare cement fabrics recognised only in the Eglwyseg Formation at Minera, between 'Eg.1' and ?'Eg.6' (e.g. s/m 1188, precise stratigraphic position unknown). Clasts are coated by a fine 150μm isopachous veneer of acicular non-ferroan calcite. This cement is pervasive, and is locally overlain by pore-filling 5 - 25μm microgranular cement, in turn overlain by a later generation of clear spar druse‡. The micro-

† Plate 16, figs B & C ‡ Plate 16, figs D & E

PLATE 16

A Fine peloid sands (MA.3.3) s/m 1246, Llwyn Hen Paro, G.R. SJ 22534610

B Fine sand and crystal silt drape over productid clast, s/m 1597, Minera. Loose block.

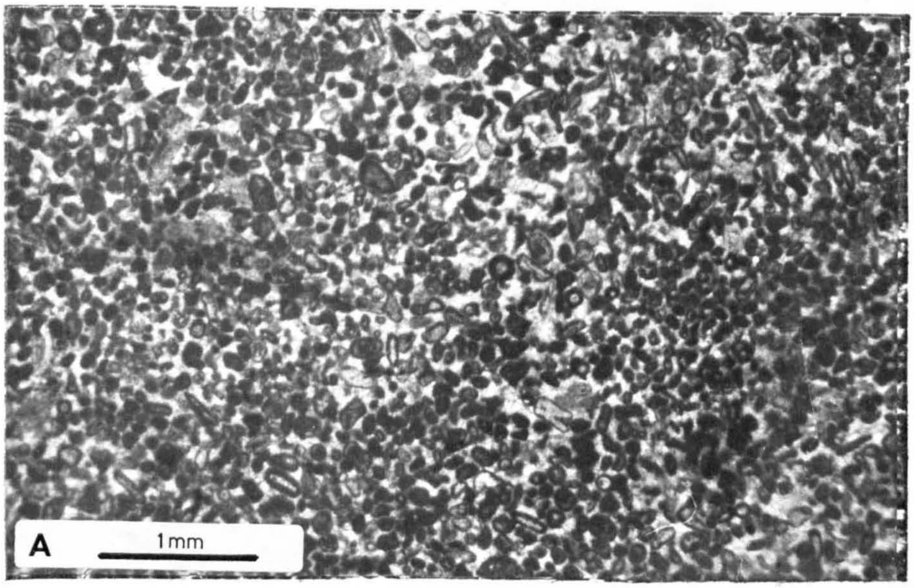
C Larger area of B. Note 'floatstone' to rudstone texture of larger clasts and fragmentary and convex-up nature of brachiopods.

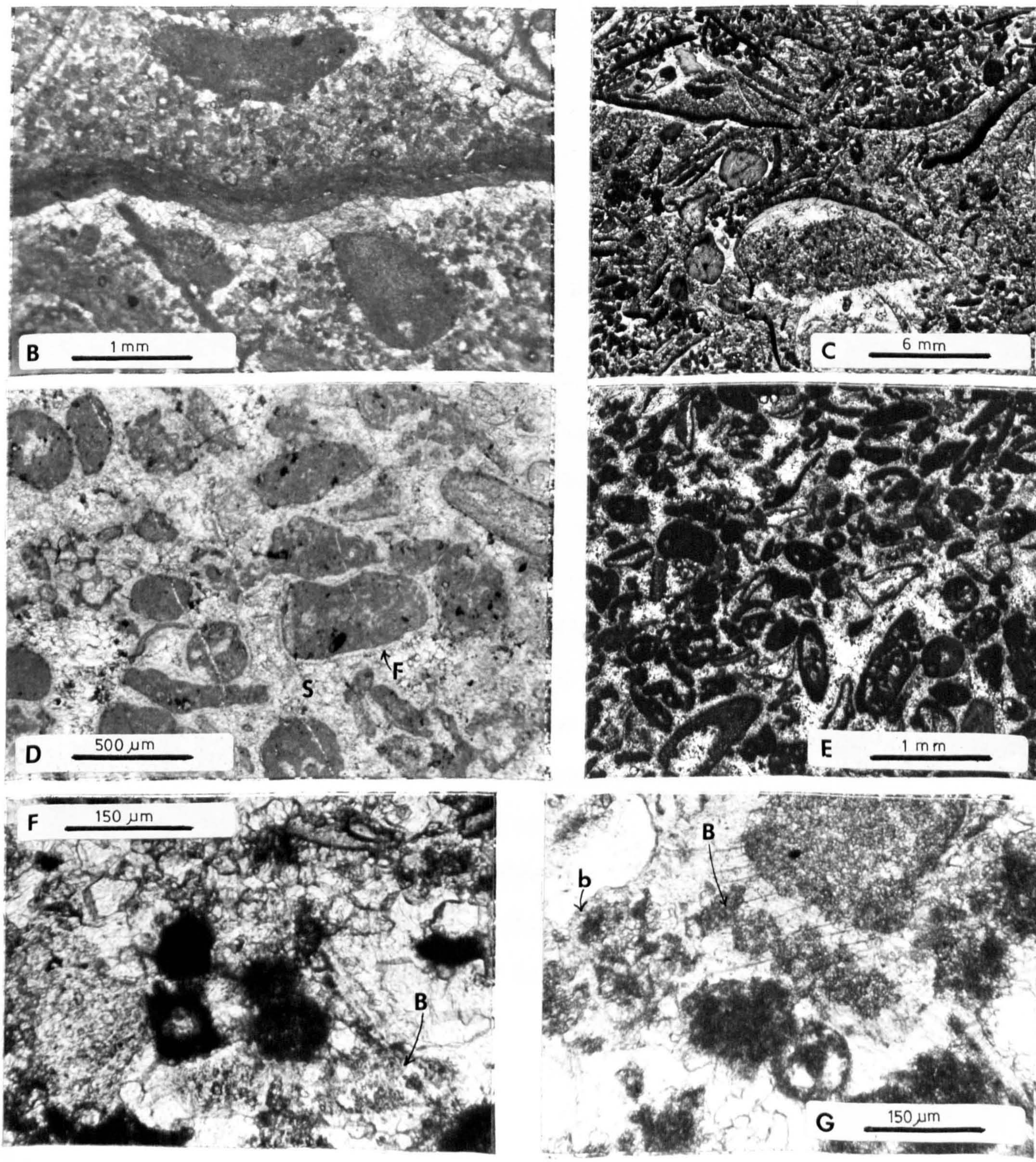
D & E Early cement fringe (F) to peloid clasts. Note crystal silt form of much of the void-filling cement. S/m 1188, Minera. Loose block. S = Spar

F Barely visible beresellid alga (B) ?*Kamaena* sp. substage diaphragm stopped down enhancing relief. s/m 034, Tynant, G. R. SJ. 22084527.

G Beresellid in common state of preservation. Note micrite-infilled a medula, and vague cortex defined by dusty inclusions, with syntaxial continuity into surrounding clear spar. ?*Kamaena* sp s/m 093, Minera, Tynant Formation, G. R. SJ25265226.

PLATE 16





granular generation does not have any alignment, and it resembles the calcite silts described below, but lacks geopetal structures (Plate 16, fig. D). The acicular fringe cements morphologically resemble Recent submarine hardground cements (e.g. Davis, 1979)

9.2.1.2. CRYSTAL SILTS AND CEMENTS. A petrographic characteristic of Eglwyseg and Llynclys Formation MA.3 sediments is the fine microgranular silt grade nature of much of the cement (5 - 50µm, comprising a few to 50% volume of the rock) imparting a poorly washed or microsparitic appearance. Much of this silt grade may be a clast fraction, either as carbonate fines or as degraded faecal peloids. This fine grained matrix probably enhances the paler colour of the sediment. The crystal silt origin of much of this 'cement' is shown in many specimens by geopetal fabrics. It occurs within bioclast cavities (towards the outer edges of the clasts) as a bottom 'sediment', with coarser spar druse overlying (e.g. dissepiment chambers and foraminifera tests), (Plate 15, fig. B), as drapes over larger bioclasts (s/m 1597ii, Minera), and as matrix, in sediment within convex-up brachiopod valves (Plate 16, fig B). These features occur in more open packed grainstones where the drapes are graded (with fines towards the top) indicating that some of the silt has been gravitationally transported.

Much silt grade 'cement' also occurs within medullae of tubular septate algae (terminology of Mamet and Roux, 1974). The cortex, enveloped in syntaxial spar, is defined by brown submicroscopic inclusions, but all grades can be seen from cortices barely visible to distinct (Plate 16, figs F&G). Bathurst (1975, fig. 55) figured similar areas of dusty inclusions, but believed them to be of fragmentary crinoid origin.

INTERPRETATION. The presence of crystal silt in bioclast cavities to the exclusion of fine bioclastic material, suggests that the silt was either able to penetrate through the sediment porosity more effectively than the bioclastic fines that the silt is a neomorphic fabric derived from original bioclastic material, or that the silt had a very local source. The second is unlikely as equally fine bioclastic debris remains well preserved, and the former would imply a sorting process in the descent of the particles. This is a possibility, but may in part be controlled by its local formation.

The transport and deposition of this silt was probably a result of vadose erosion, as described in section 6. 4. 6, in which case much of the silt itself may be primarily a vadose cement. In less 'open' grain support frameworks there is no obvious draping effects, and the silts pervade the sediment. It is in these more typical sediments that the origin of the 'silt' remains speculative, and it is in these sediments that '<u>structure grummeleuse</u>' (Bathurst 1975) is most readily discernable.

9.2.2. Pseudobreccia Mottling in Asbian sediments.

Colour mottling from a millimetre to tens of centimetres scale is common throughout the MA.3 microfacies of the Eglwyseg and Llynclys Formations, and is rarely present in Trefor Formation sediments. Characteristically the mottling varies from strong colour contrasts (always darker mottles) to only vaguely discernible ones. The mottles may be tightly packed and stylolitically interpenetrating, or very scattered. Interpretations of mottle formation fall into two schools, either a bioturbation origin, with mottles defining burrow systems, or a diagenetic selective recrystallisation product. Two basic forms of colour mottle are recognised in this study.

9.2.2.1. CEMENT-RICH MOTTLES. This form encompasses the most common and typical mottle type. They occur in a transition zone between well bedded darker MA.3 sediments and the paler massive MA.3 of minor cycle upper phases. This zone may vary from tens of centimetres to metres. They also occur scattered throught the massive MA.3 units defining vague to prominent layering, by variations in shape, size and density†. They rarely exceed 8cm diameter, but vary from rounded to irregular margins, often diffuse, and from subspherical to irregularly elongate. Stylolites tend to circumscribe their margins, and preferentially remove intermottle sediment by pressure solution. Towards the outer margin of these mottles there is occasionally a slight dark rim, although this varies from mottle to mottle, and is usually impersistent around their perimetres (e.g. s/m 1238)

The mottles are occasionally partly filled by large bioclastic fragments. Mottles define the burrow structures of Plate 17, fig.C.

Petrographically these mottles appear to contain similar clasts in similar proportions to the inter mottle matrix. However, characteristically these mottles contain more spar cement than the matrix, which often appears well compacted (e.g. s/m 1092 Sychtyn Member, Nant Mawr Quarry). This compact nature of the paler matrix is visible on both polished surfaces and peels. The greater the difference in compaction between mottle and matrix, the more pronounced the colour distinction. Clasts close to mottles tend to be aligned sub-parallel to the mottle margins. Rarely do clasts penetrate from mottle to matrix, primarily (?) due to their stylol-

† Fig. 59

FIG 58

EXAMPLES OF MOTTLE DEVELOPMENT

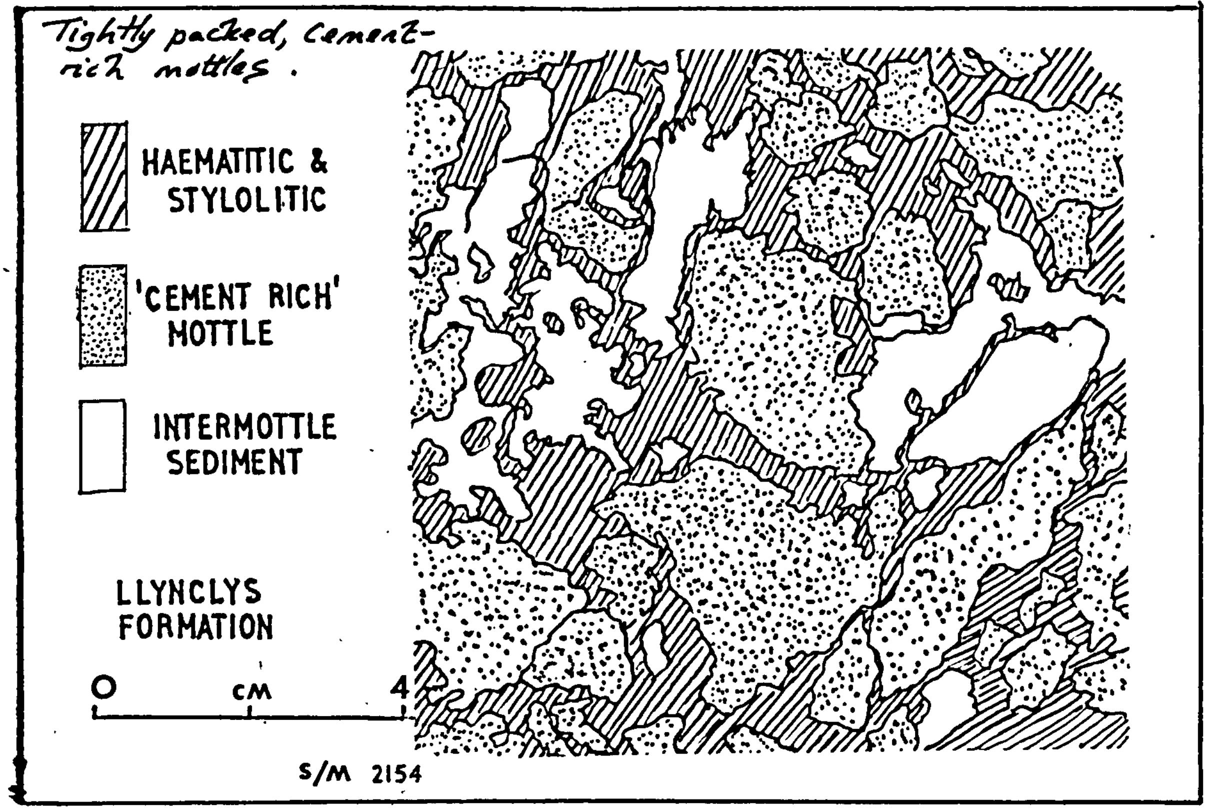
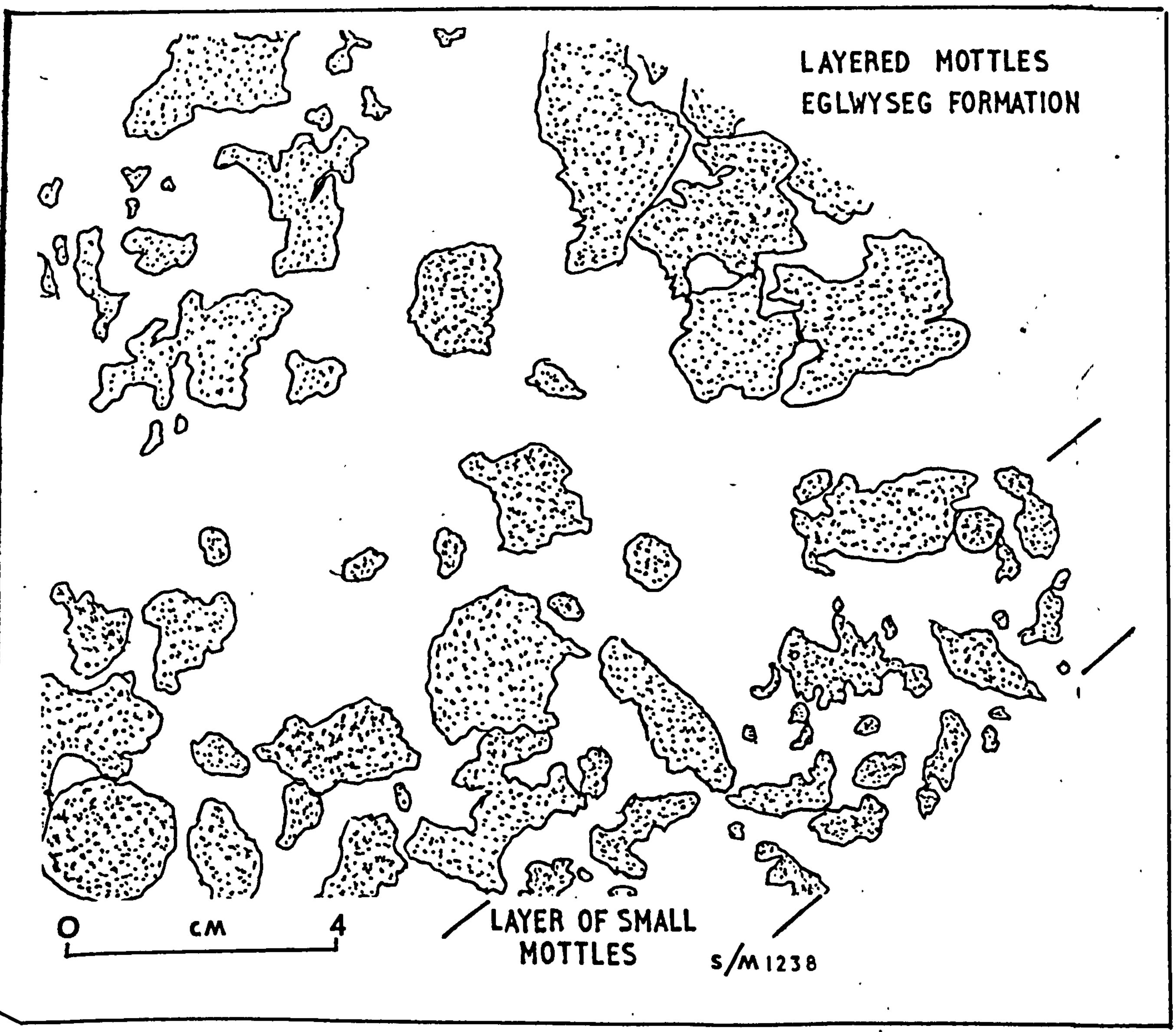


FIG 59



itio fringes. Stylolitisation may go to the extreme, producing a stylonodular texture in which each stylonodule defines a single mottle, with only minor quantities of paler sediment between (e.g. s/m. 2154, fig.58).

Irregular darker mottles often occur within or surrounding coral bioclasts or colonies in sediments of paler MA. 3 lithology (Fig. 61). Corallites of fasciculate colonies may be readily defined by their incorporation within the mottle whilst the colony itself is apparently entire or little fragmented. Vague colour mottling also occurs within cerioid lithostrotiontid colonies. These bioclast-related mottles occur in close association with the irregular mottles previously described, and are, therefore, believed to have similar origins.

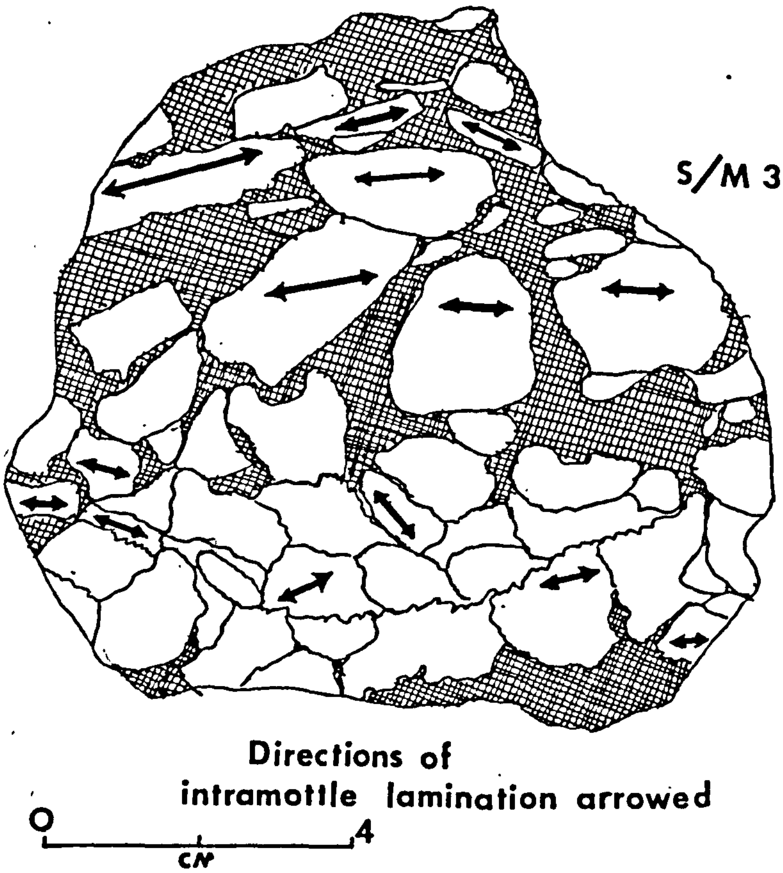
Other mottle forms occur on this cement rich theme. S/m 1503 has sub laminoid darker irregular mottles (2cm thick) related to areas of coarser 'well washed' bioclastic grainstone (dominantly crinoidal), with coarse calcite spar cement, in comparison to the peloidal, compacted intermottle matrix. At Trefor Rocks (G. R. SJ 22854333, 'Eg. 3, 3a') a centimetre scale dark-light laminated alga peloid grainstone (MA.3.1) grades laterally into irregular mottle horizons, through all stages from interconnected mottles to scattered and only vaguely related to the layering, over a distance of two metres (Plate 17, fig. A).

In s/m 464 ('Eg.1.2', Trefor Rocks) and s/m 1461 (Sychtyn M'br, Craig y Rhiw) elongate ellipsoid mottles occur, with smooth margins, but with internal lamination extending through the mottle to the surrounding compacted intermottle matrix.

The largest mottles found occur within MA.3 microfacies of the Sychtyn Member. They are dark, with smooth ellipsoid form, up

FIG 60

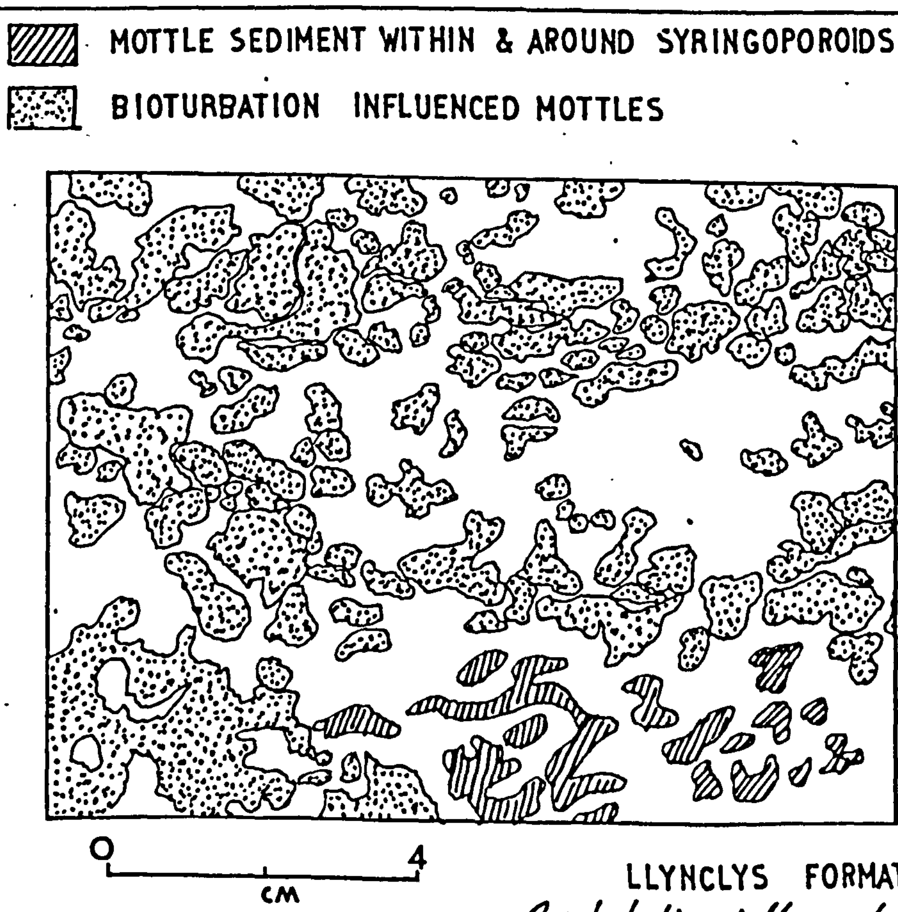
A



CEMENT RICH LAMINATED MOTTLES , EGLWYSEG FORMATION

B

FIG 61



LLYNCLYS FORMATION

Bioturbation influenced mottles.

to 15cm long and retain a distinct grainstone fabric and are defined by stylolites that separate them from the surrounding compacted intermottle matrix.† S/m 320 (?'Eg. 5. 6' Trefor Rocks (loose scree)) shows irregularly shaped, 3cm diameter cement rich mottles, densely packed by stylolitic compaction, but some mottles show internal millimetre sedimentary lamination, and (Fig. 60) is at a different angle to the bedding of the rock. A degree of rotation must have occurred to achieve this, probably through the pressure solution process.

<u>INTERPRETATION</u>. Mottle sizes and shapes are variable, but being of a constant form and density along individual layers indicates that their shape is more sediment controlled than by external factors. The lack of varied clast lithologies, micritised rims, and the rare persistence of fabrics between mottle and matrix indicate that the mottles are not intraclasts.

There is neither solid evidence to argue for or against a common formation for all the mottle forms described in this section although their association and external similarities suggest related formation mechanisms. The presence of lamination within the mottles, along with their formation within and around bioclasts indicates that for at least some bioturbation was not involved as an initial process of mottle formation. Bioturbation of intermottle matrix may have occurred in some. Homogeneous sediments do not provide any information on this aspect, although their thorough homogeneity suggests pervasive bioturbation of the unlithified sediment.

Rose and Dunham (1977, p.38) also recognised a "higher content of recrystallised matrix " within the mottle, and a flow texture

† Plate 17, fig. B

in the intermottle sediment, imparted by compaction and microstylolitisation. This evidence, corroborated above, suggests that local intra-mottle cementation occurred prior to compaction, was often layer-controlled, but was of a different form to any previous cementation within the intermottle sediment.† According to Meyers (1977) compaction occurs on burial to metres or tens of metres. Bathurst (1975, p.465) notes that grain to grain pressure solution cannot take place after precipitation of a second pore filling cement generation although it may happen after a first generation partial cementation. There is no evidence of the latter in intermottle cements (apart from 'crystal silts' (?)), which indicates that little if any cementation had occurred within intermottle matrices prior to compaction. Therefore the mottle cements are early diagenetic events that prevented grain-grain compaction and pressure solution. The origin of this cement, (presumably nucleating around incipient mottle centres before burial past tens of metres) remains unclear.

9.2.2.2. <u>MATRIX - RICH MOTTLES.</u> In more stylonodular and argillaceous sediments neomorphic calcite growth provides a mottle/nodular texture, described in section 9.1.2.

Matrix-rich trace fossils (e.g. chondrites sp., Plate 11, fig. G) and irregular burrow forms (9.1.4.1., s/m 1246) produce mottling in vertical profile.

---

† <u>cf.</u> section 10.5.1 .

PLATE 17

A Layers of mottles and continuous ?primary layers of similar colouration and petrographic form to the mottles (dark). Eg. 2a. 3, Trefor Rocks. G.R. SJ22834335

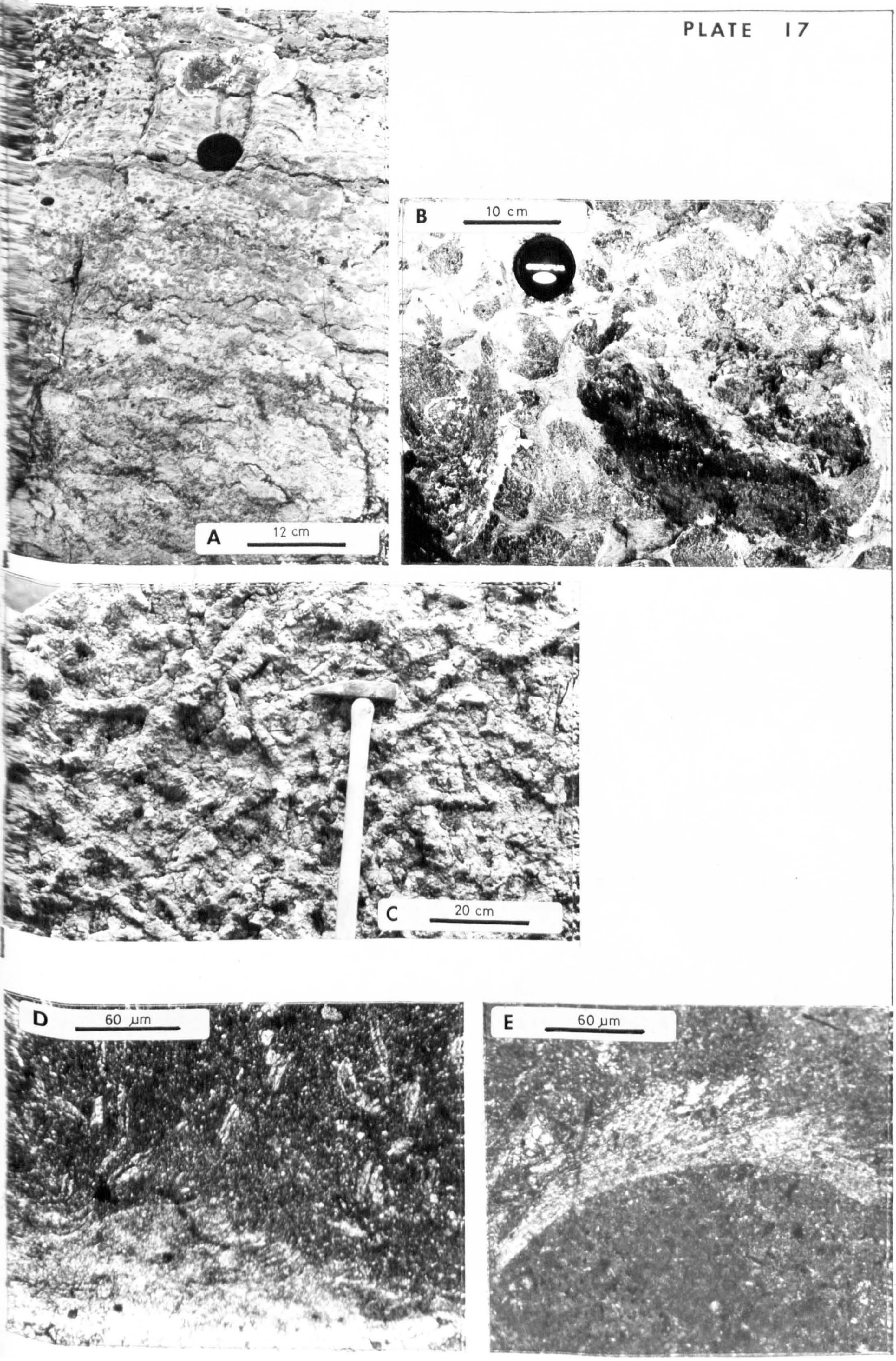
B Large subspherical mottles with well defined edges. Cement rich mottles in compacted and microstylolitic matrix. Loose slab, Minera (?from upper strata of the Eglwyseg Formation)

C Distinct bioturbation. Burrow forms weathering out in stylonodular and mottled MA.3 sediments. Burrows conform to cement-rich mottles. Loose slab. Mid-upper Eglwyseg Formation, Llwyn Hen Parc. (similar to Garwood's (1913) 'Stick Bed'

D Edge of a syringoporoid corallite showing <u>in situ</u> degradation with prisms of the lamellar walls 'spoiling' off. (?a bioturbation effect) s/m 174, World's End, Eg.1.2. peel G. R. SJ 2334812

E Clast of <u>Composita</u> showing <u>in situ</u> degradation, with prisms of the impunctate microstructure spoiling off (cf D). s/m 2150, Sychtyn Member, Nant Mawr Quarry G.R. SJ 25236503.

PLATE 17



9.3. PALAEOENVIRONMENT OF THE EGLWYSEG FORMATION.

9.3.1. A Model for Sedimentation of MA.3 Microfacies.

9.3.1.1. 'MATURITY'. Sediments of MA.3 microfacies are here classified according to their degree of 'maturity'. Maturity in this context represents the degree of alteration (contemporaneous) of the biogenic components of the sediment. A very immature sediment is one composed totally of unfragmented bioclastic material that has suffered negligible contemporary corrosion or alteration, either biochemical, biomechanical, chemical or mechanical. A highly mature sediment comprises well sorted peloid clasts (micritised bioclastic debris).

Most MA.3 sediments in this study fall between these extremes, either due to admixing of sediments of different degrees of maturity, or less rarely, uniform but only partial development of these sorting, degradation and micritisation processes.

9.3.1.2. FACTORS CONTROLLING DEGREES OF MATURITY.

1. Algal Micritisation. Bathurst (1975, p.317)noted that Bahamian grapestones are intensely micritised by the filling of vacated algal bores with micritic aragonite, to such an extent that unaltered skeletal clasts are very rare. Micritisation is common throughout MA.3 sediments, identified by pale rims to clasts in polished section, and is mostly endolithic (Bathurst, 1966) and rarely exolithic† (Kobluk and Risk, 1977). Whether micritisation rate depends on clast type is unknown, but all bioclast types do succumb to it. To effect algal-induced micritisation the sediment particle must receive light, and is probably enhanced with water movement over the clast

† e.g. Frontispiece.

surface. Algal induced micritisation can, therefore, occur only within the very surficial sediment layer.

2. Bioclast production. The type and rate of production of bioclasts will determine, in part, their relative proportions in the final sediment. Higher rates of sedimentation, induced by higher production rates will reduce the length of time clasts are susceptible to micritisation, mechanical fragmentation and degradation.

3. Skeletal degradation. Different organisms produce bioclasts of differing microstructures, each with different fragmentation and degradation characteristics. Some readily fragment by mechanical attrition, some remain as resilient sand grains, whilst others degrade *in situ* by biochemical corrosion (decomposition) of binding organic materials. Syringoporoids have been noted contributing prisms of wall microstructure to the sediment between corallites† (s/m· 174). *Composita* sp. and *Camarotoechia* (Plate 17, fig. E) also contribute discrete calcite prisms to the sediment, readily observed in floatstones and rudstones.

4. Mechanical disintegration and transport. It is primarily these two related processes that account for the non-uniform character of MA.3 microfacies. Mechanical abrasion provides a spectrum of clast sizes that are susceptible to biochemical and biomechanical alteration. Micritisation of clasts tends to strengthen them reducing chances of further fragmentation (Bathurst, 1975). Transport of clasts, either as saltating bedload, or in suspension admixes clasts of different generations (degrees of maturity). Thus MA.3 microfacies (especially MA.3.1) suggest a high degree of sediment movement.

---

† Plate 17, fig. D

This correlates with the probability of deposition above or near to fair-weather wave-base. MA.3.2 microfacies reflect either reduced biochemical alteration, increased sedimentation rate, or little sediment movement admixing the clasts. Conversely MA.3.3 sediments reflect either reduced sedimentation rates, or high rates of biochemical alteration, but also suggest that little sediment admixing occurred.

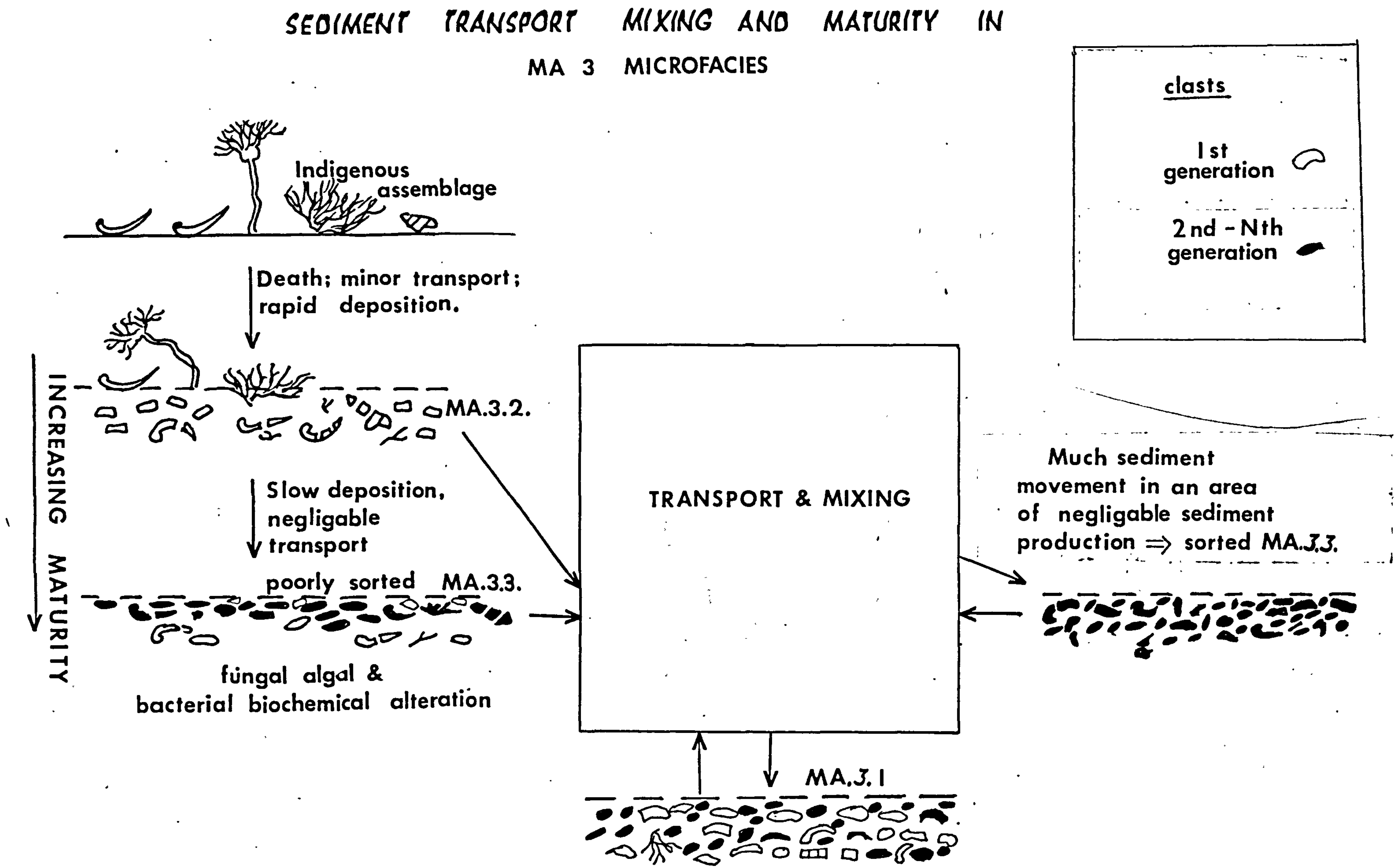
Fig.62 illustrates the inter-relationship between these three microfacies.

9.3.2. Some aspects of Biostratinomy.

Faunal horizons within the Eglwyseg and Llynclys Formations are readily divided into two basic forms:

1. Those associated with the massive MA.3 upper phase, in which thin horizons contain large fragmentary and overturned macrofossil elements (coral-brachiopod)
2. Those associated with basal phases that comprise in-growth coral-brachiopod assemblages, rarely meriting the term 'biostrome'. The stylonodular phase at the base of cycle 'Eg.8. 8a', is noteable, comprising in growth Lithostrotion maccoyanum, overgrown by Syringopora sp, and associated with in-growth colonies of L. junceum (see fig. 55). Shell beds of Gigantoproductus of maximus also occur in similar basal stylonodular phases of higher cycles (e.g. 'Eg.9a.10', Dinbren Uchaf), but S.D.D. values (see fig. 53) do not go below unity (?).

In the bedded MA.3 units of transgressive phases, coral colonies are mostly fragmented, and many overturned (indicating a degree of transportation, although many colonies, appear fragmented



Model relating composition of MA.3 sediments to in situ degradation / alteration and mixing through transport of clasts of different generations (degrees of maturity).

Fig. 62

nearly in situ, with clasts moving only centimetres from the main colonial mass (Plate 2 fig. E), confined to impersistent layers.

The degree of fragmentation and clast dispersion is even more marked in the massive bedded MA.3 phases. Fasciculate lithostrotiontid colonies are often represented by a lens of separated corallite debris, and many solitary corals have marginaria highly eroded. These layers probably attest to frequent storms moving bottom sediment and biota.

9.3.3. Summary of Eglwyseg and Llynclys Formation Palaeoenvironment.

Somerville (1977, 1979a) described the deposition of Eglwyseg Formation cycles in terms of a regressive progradation of facies overlying a thin basal transgressive unit.

Basal transgressive phases represent deposition beneath fair weather wave base, associated with a ?rapid transgression. They comprise both stylonodular and thin bedded MA.2 and MA.3 sediments. Clastic influences are variable, but clays are more significant than in the regressive phases. The MA.2 and MA.3 sediments are tubular-septate algae rich, and support a stenohaline coral-brachiopod assemblage.

Regressive phase sedimentation developed massive MA.3 units, apparently deposited above fair-weather wave base. Sediment movement was common, and storms periodically ravaged the substrate, but high current energy bedforms are rare. Early cementation of these sediments was rare, but much of the silt grade cement fraction may have been 'deposited' during periods of vadose erosion, associated with emergence in the latter regressive stages. The pseudobreccia mottling prevalent in these regressive phases is a diagenetic effect, associated with a cementation event before deep burial, at times associated with bioturbation phenomena.

10. MICROFACIES ANALYSIS OF THE TREFOR FORMATION

(Basal transgressive phase of Sandy Passage Form'n included here)

The Brigantian Trefor Formation represents a subtle change in the style of sedimentary minor cyclicity, with the development of thick basal phases of thin bedded argillaceous biomicrites (Lithofacies L.3), dominated by sediments of the argillaceous algal packstone microfacies association (MA.2) and thin but persistent massive bedded pale biosparites and micrites of L.1 and L.5 (MA.3 and MA.1 microfacies associations) of minor cycle regressive phases.

Many important sediment contributors in Asbian palaeoenvironments are lacking or are much reduced in the Brigantian, e.g. beresellid and <u>Koninckopora</u> dasycladacean algae. Other algae do, however, assume a significant contributing role <u>in lieu</u> of these Asbian elements, paramount of which is the dasycladacean <u>Coelosporella</u>, a rare component of Asbian sediments.

10.1. TRANSGRESSIVE PHASE SEDIMENTS OF TREFOR FORMATION MINOR CYCLES.

<u>Microfacies represented:</u>

Argillaceous algal packstone microfacies association.

MA.2.1. Algal packstone to wackestone.

MA.2.2. Bioclastic packstone to wackestone.

10.1.1. <u>Distribution and Field Character of MA.2.</u>

Basal carbonate phases of minor cycles are thin, wavy or irregularly bedded (5 to 50cm) dark grey to black argillaceous packstones. They are often associated with thick argillite phases and have centimetre argillaceous partings, and are rarely internally parallel-laminated.. They support a diverse stenohaline fauna dominated by corals and brachiopods.

10.1.2. Petrography of MA.2.1. Alga packstone to wackestone

This microfacies resembles MA.2.1 of Asbian sediments, except that beresellid and tubular/septate algae are minor components. The dasyclad Coelosporella jonesi is the most significant contributor. Matrix is similar to associated MA.2.2. Girvanella nodules (up to 1cm in diameter) also occur within this microfacies towards the base of the Trefor Formation (s/m 722, Craignant, Plate 18 fig. B)

10.1.3. Petrography of MA.2.2. Bioclastic packstone to wackestone (Plate 18, fig. E)

This is a petrographically diverse microfacies. The matrix of bioclastic silts and lime/argillaceous muds is often microsparitised or pseudosparitised. Locally significant contributors include crinoids, brachiopods, gastropods, Saccamminopsis sp, textularid foraminiferas and fenestellid bryozoans.

Beresellid and other tubular septate algae, as with Brigantian MA.2.1, are present in very minor proportions. Calcispheres are also minor but ubiquitous components.

Zoophycos and irregular subhorizontal burrows commonly pervade the sediment.

Micritisation of clasts is neglibible, although they commonly show 10 – 20 µm diameter microborings on their surface.

Heterocorals (Hexaphyllia sp) are also rare components of this Brigantian microfacies.

10.1.4. Standardised Petrographic Diversities of MA.2 sediments.

Shown in table 5 , the standardised petrographic diversities of Brigantian MA.2 sediments are the highest recorded in the study-succession. Total diversities of 23 $\pm$ 3 are a result of high macrofaunal diversities, reflected in the diverse faunal assemblages within the sediments (see appendix V). Microfaunal diversities are also slightly higher than other microfacies associations in the Brigantian and Asbian (9 $\pm$ 1), due to the presence of forms including Saccamminopsis and tetrataxids. Algal contributions are uniform due primarily to the ubiquitous (if not prolific) calcispheres.

10.2. REGRESSIVE PHASE SEDIMENTS OF TREFOR FORMATION MINOR CYCLES.

Principle Microfacies recognised:

Alga-peloid grainstone microfacies association (MA.3).
- MA.3.1. Alga-peloid grainstone to packstone
- MA.3.2. Bioclast grainstone
- MA.3.3. Peloid grainstone.

Calcisphere wackestone microfacies association (MA.1)
- MA.1.5. Coelosporella mudstone to wackestone.

10.2.1. Distribution and Field Character of MA.3.

As with MA.3 sediments within the Asbian, this microfacies association occurs as the major component of upper regressive phases of typical minor cycles, as paler (medium to light grey) massive-bedded units of Lithofacies L.1.

Pseudobreccia mottling occurs within this association. Biostromes are less common than within MA.2 sediments. MA.3 units are $\leqslant$ 8m thick, and often have emergence phenomena on their upper surface but at times they grade upwards into developments of MA.1.

Sediments of packstone texture are relatively common within these regressive phases, although this is enhanced by matrix-volume reduction through compaction.

PLATE 18 <u>Trefor Formation</u>

A Coelosporella wackestone (MA.1.5). s/m 1205, Pant Hir Quarry, G. R. SJ 238280

B Exolithic micrite rim + Girvanella oncolitic coating (C) on a pseudopunctate brachiopod clast. s/m 722, Craignant, G. R. SJ 254348.

C Alga peloid grainstone-packstone MA.3.1. Note poorly washed nature, and the peloid-grainstone with micrite infilled Coelosporella utricles giving the illusion of rounded peloids. s/m 215. Trefor Rocks.

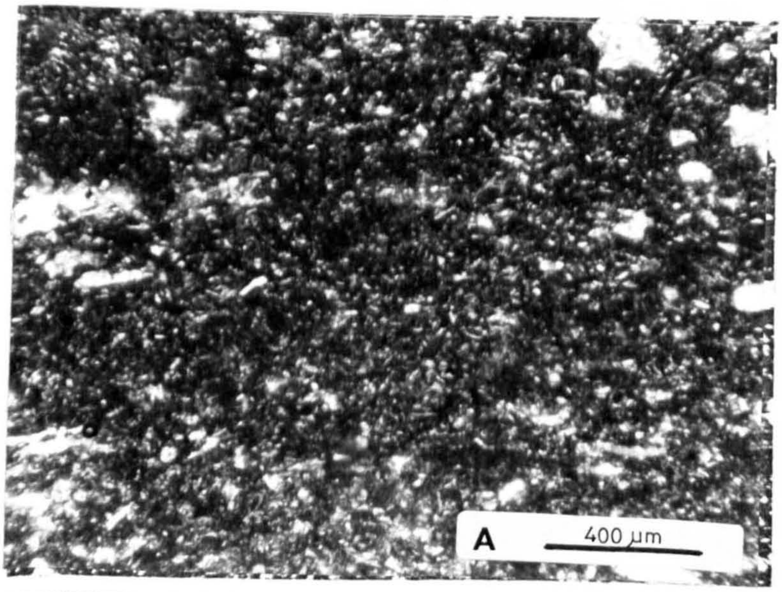
D Intermottle matrix with partly collapsed moulds of Coelosporella, (C) recognised by micritised 'peloids' of utricle infills (U). S/m 070, Trefor Rocks. G. R. SJ 23334330.

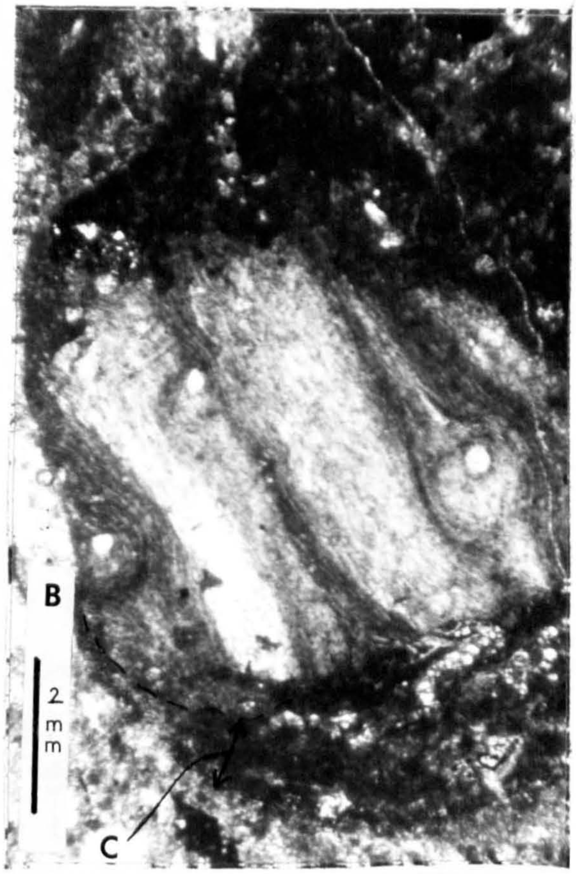
E Bioclast packstone MA.2.2. S/m 421, Froncysyllte. Stained peel.

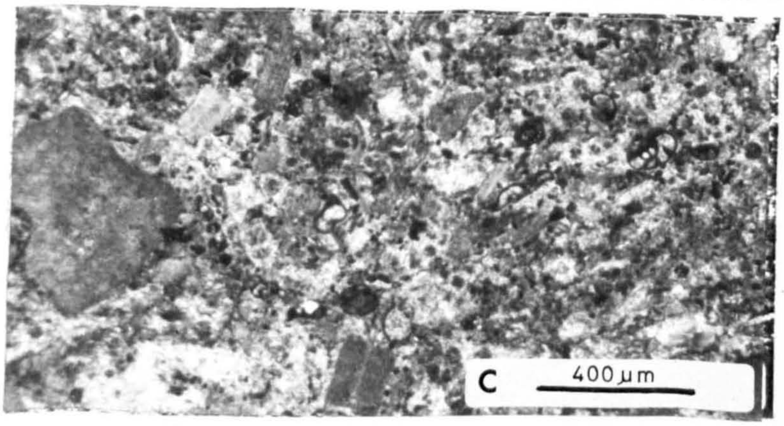
F Coelosporella-rich (MA.3.1) mottle within compacted and microstylolitic-sediment, S/M 1360 . Llawnt. Stained acetate peel

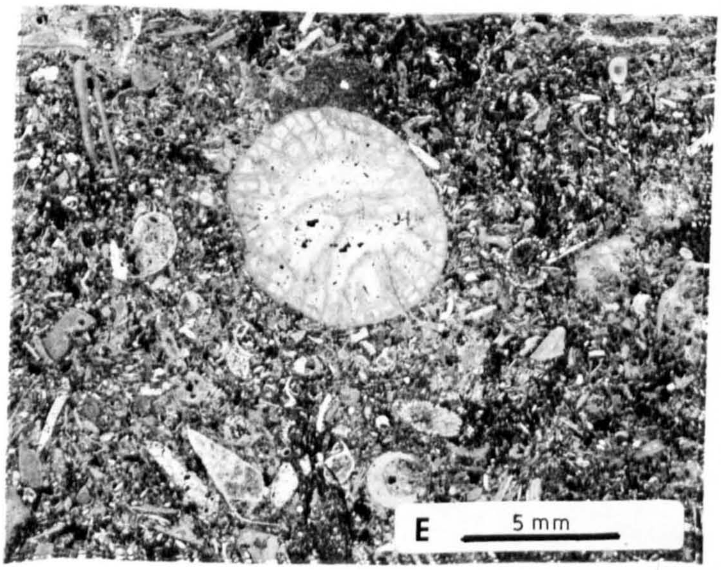
G Minor 'bioherm' of Diphyphyllum furcatum, Trefor Rocks, G. R. SJ 23054334

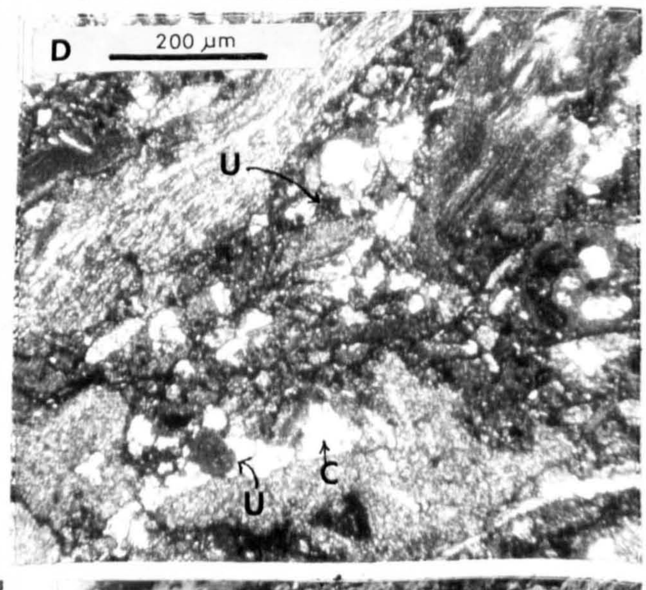
PLATE 18

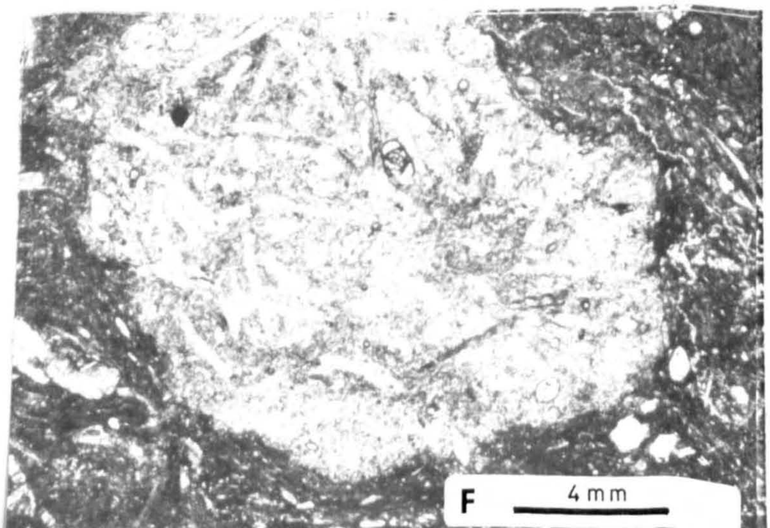


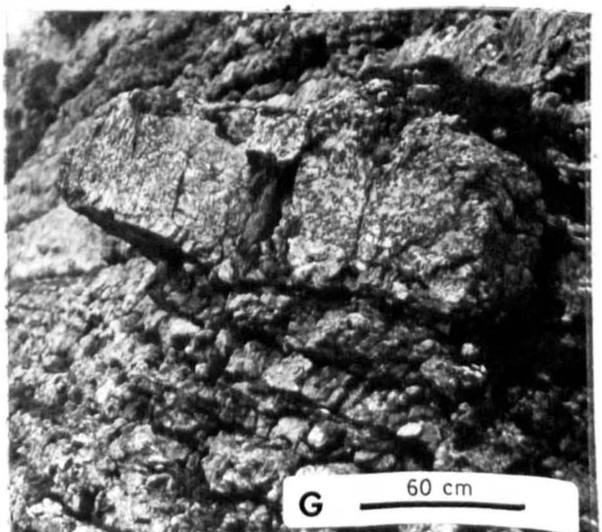












10.2.2. Petrography of MA.3.1. Alga Peloid Grainstone/Packstone (Plate 18, fig. C)

Whereas the tubular-septate beresellids and *Koninckopora* were important contributors to MA.3 sediments of the Asbian, *Coelosporella* (a dasyclad) is the Brigantian equivalent (Wood,1940).

At times accounting for up to 50% by volume of the sediment, *Coelosporella* clasts vary from large to small fragments of the thallus. Utricles of *Coelosporella*, infilled with micrite (of beresellid medulae), superficially resemble peloids in section. Calcispheres are minor components.

These sediments commonly have a ' poorly washed' appearance, with bioclast fines, mud grade carbonate, and silt grade cement admixed as matrix. 'Well washed' examples are exceptional. As such this microfacies grades into MA.1.5 and MA.2.2. Macrofossil clasts, primarily brachiopod crinoid and coral, are commonly micritised, whilst algal clasts are not. The sediment is mostly a poorly sorted 250 µm fine carbonate sand.

Irregular burrow forms, up to 1cm diameter, occur as a more open packed framework, but *Zoophycos* is rare compared to transgressive phase MA.2 sediments.

10.2.3. Petrography of MA.3.2. Bioclast grainstone.

This is a minor microfacies present within MA.3 dominated units of the Brigantian. As with MA.3.2 Asbian sediments, it comprises $\geqslant$20% recognisable macrofossil clasts. Brachiopods and crinoids are the most significant contributors.

With increasing mud matrix, MA.3.2 grades into MA.2.2, although as with MA.3.1 it is commonly 'poorly washed'.

10.2.4. Petrography of MA.3.3.

Similar in composition to Asbian MA.3.3., this minor microfacies within the massive-bedded pale regressive phases, is characterised by > 50% peloids. Mostly well sorted ≤ 500µm sands, this microfacies occurs in units 10's of centimetres thick, with Coelosporella as the major indigenous bioclast. Internal lamination is often well defined within this microfacies, reflecting periodic transport/deposition and a lack of bioturbation.

Micritised clasts, and the presence of MA.3.3, are noteable immediately underlying calcrete crusts (e.g. s/m 1362, Llawnt 'Tf.3').

10.2.5. Petrographic diversities of MA.3 Sediments.

Standardised petrographic diversities of Brigantian MA.3 sediments (table 5) are very similar to late Asbian MA.3 (total 18 ± 3), with equal contributions from macrofauna microfauna and microflora. This indicates negligible environmental restriction (sensu Wagner and Togt (1973)), although the macrofauna (?essentially stenohaline elements) is only two thirds of stratigraphically adjacent argillaceous MA.2 counterparts.

10.2.6. Distribution and Field Character of MA.1.

In character, this microfacies association is a pale to dark grey "porcellanous micrite", massively bedded, and occurring beneath emergence surfaces (palaeokarsts calcretes and sutured discontinuity surfaces) as laterally impersistent regressive phases of some Trefor Formation minor cycles. It is developed primarily at the top of the fifth minor cycle ('Tf. 5. 6') between Trefor Rocks and Llawnt (fig. 30) the 6th, 7th minor cycle (beneath 'Tf. 7'&'Tf.8')

along Eglwyseg escarpment and as a minor development underlying 'Tf.2', 'Tf. 3' and 'Tf.4' exposed at Trefor Rocks (Chart D.).

10.2.7. Petrography of MA.1.5. Coelosporella wackestone to mudstone (Plate 18 fig. A)

This is the only microfacies of MA.1 suite recognised within the Trefor Formation. Calcispheres, the dominant bioclast component of Asbian MA.1 microfacies are minor contributors, whilst Coelosporella is locally significant. Peloids are dominant components, and include smooth eliptical forms < 100μm diameter of probable faecal origin. Foraminifera, crinoid and ostracod fragments are ubiquitous but minor contributors. The matrix comprises mud and silt-grade carbonate locally neomorphosed to a microspar/pseudospar texture. Lamination is vague to indistinct, especially reflecting variation in peloid content.

With increased bioclast contribution and clast support MA.1 grades into MA.3.

Tubular fenestrae (section 6.4.1.3) with internal sediments are rare, but other fenestral fabrics are absent. Bioturbation includes Chondrites sp, and irregular sub horizontal burrow forms ≤5mm. diameter. Indigenous macrofauna is rare. Occasional colonies of Lithostrotion pauciradiale (in 'Tf. 7. 8' MA.1 phase) are overturned, attesting to at least local disturbance

10.3. PALAEOENVIRONMENT INTERPRETATION

10.3.1. M.A.2 Microfacies.

Apparently deposited below a 'fair weather wave base' sediments of MA.2 reflect normal marine deposition, with a substrate supporting a diverse macrofaunal assemblage. The absence of abundant algal material may reflect the lack of available contributors. Coelosporella is the most significant alga, and being a dasycladacean, suggests shallow low energy bottom conditions (Wray 1977).

Analogies may be drawn from the more (axial) offshore regions of the Arabian Gulf (Wagner and Togt, 1973) and the "Basal Sheet" sediments of Shark Bay (Hagan and Logan, 1974, see section 7.2.8.)

10.3.2. MA.3 Microfacies.

Brigantian MA.3 depositing palaeoenvironments are broadly analogous to their Asbian counterparts as shown by similarities in sediment type and position within the minor cycles. Deposition above or close to a normal wave base is suggested by the mixing of clast generations and the 'washed' or 'partly washed' nature of the sediment texture. A relatively diverse coral-brachiopod fauna attests to stenohaline conditions, with coral biostromes locally developed (see on). Minor cycle boundary emergence events commonly occur on top of MA.3 microfacies, lacking development of MA.1 sediments, and, therefore, as with Upper Asbian strata, are "grainy" minor cycles (Walker, 1979), although evidence of shoreface environments is lacking, suggestive of relatively rapid regression from a shallow subtidal situation leading to emergence.

10.3.3. M.A.1 Microfacies.

There is a marked absence of sedimentary fabrics and structures that may indicate a persistently tidal-emergent situation for these sediments. Their presence as lime- mud rich microfacies towards the top of shoaling minor cycles analogous to Mesothem D5a, is the prime peritidal indicator. Fenestral fabrics are lacking apart from tubular forms that in Asbian sediments, tend to be associated with basal fenestral phases of progradational units. Their upper surfaces are often palaeokarstic and rarely calcretised (e.g. 'Tf.7' Trefor Rocks). An intertidal sutured discontinuity surface (section 11.1) occurs below 'Tf.3' at Trefor Rocks. Lack of algal films (and cryptalgal sediments) may have, in part, contributed to the absence of fenestral fabrics. Certainly, neither penecontemporary dolomites, early cements nor pisolitic (coated) grains have been observed.

A shallow subtidal environment is, therefore, probable for most of the MA.1 sediments of the Trefor Formation.

This MA.1.5 microfacies is also laterally equivalent to, and overlies, lime-mud rich MA.3 microfacies of shallow subtidal origin.

These factors imply that the sediments were deposited within a low energy progradational 'peritidal' environment. A relatively rapid progradation/regression and low tidal range may account for the lack of intertidal features.

Eustatic regression probably retarded progradation, especially if the regression was relatively rapid. This may account for the lateral impersistence of MA.1 microfacies, and their palaeogeographic location towards depocentres, where eustatic regression may, in part, be counteracted by subsidence.

10.4. SOME ASPECTS OF FAUNAL ASSOCIATIONS WITHIN THE TREFOR FORMATION SUBTIDAL SEDIMENT SUITE.

Bioherms and Biostromes.

Organic growths in the Brigantian are primarily laterally

COMPOSITION OF BIOSTROME 'Tf 1a.2, TREFOR ROCKS, G. R. SJ 23004339

Component	No.colonies (%); N=60 A	total area of colonies (%)	av. colony area (cm^2) N=60	% in growth orientation	No.colonies (%); N=54 B	Total No. colonies(%) A+B; N=119
Diphyphyllum sp + Lithostrotion martini	5	6	570	100	11	8
L. irregulare + L. paucirad L. pauciradiale	23	18	330	80	6	15
L. junceum	3	1	90	50	6	4
L. decipiens	35	45	540	70	60	46
Lonsdaleia floriformis	6	9	600	50	7	7
Palaeosmilia regia	8	10	560	80	0	4
Syngopora cf ramulosa	8	4	230	60	4	6
S. cf geniculata	12	7	230	80	7	10

Biostrome A: Total colony area $27400cm^2$

Total quadrat area 37m x 130cm

Biostrome B: Same biostrome, 400m south of SJ 23004339,

Table 7

persistent thin biostromal units lacking a skeletal framework. These biostromes are either coral or brachiopod dominated.

The natural exposures along Trefor Rocks have provided an ideal site for a semiquantative investigation of these units using line quadrats and lateral logging (Ager, 1963).

Bioherms are limited to metre-scale aggregations of coral colonies, within Lithofacies L.1 and L.5.

10.4.1. Monospecific Coral Bioherms.

Within the basal thin bedded phase of 'Tf.1.2.' along Trefor Rocks (G.R. SJ.23054334) is a biohermal growth of dominated by lithostrotiontid Diphyphyllum furcatum (Plate 18 fig.G.). This growth, 7m by 60cm tall, has a basal zone of flat-lying (uncrushed) corallites. Within the upper 40cm of the mass, the corallites are subvertical and in growth position. The matrix to these colonies (number of discrete colonies not appreciable) is bioclastic wacke-stone (MA.2.2) of Lithofacies L.2 and supports a fauna including Productus sp, Gigantella sp, Brachythyris sp and Megachonetes sp and an ?encrusting colony of Syringopora. sp.

On the bioherm top there is also a veneer, a few centimetres thick, of horizontal, and slightly crushed corallites (indicating a degree of compaction) that are aligned at 070°, indicating currents may have disturbed and reworked the bioherm surfaces prior to burial. The bioherm is laterally equivalent to two 30cm beds separated by a bedding plane.

For bioherm growth to have negligible relief above the sediment water interface, growth rates would have had to have been comparable to sedimentation rates, that at maximum, in comparison to Recent environments were 120cm /1 000y. (Chapter 12).

According to Johnson and Nudds (1975) comparable growth rates of Lithostrotion martini (fasciculate lithostrotiontid) were between 3 and 5mm per month. It is, therefore, probable that this bioherm had a high relief if growth was a continuous process.

10.4.2. Multispecific Bioherms. (including those developed within Sandy Passage Fmn.)

Minor multispecific coral bioherms of metre size occur throughout the L. Brigantian. They have a central colony support structure, and may grade laterally into biostromes.

Exposure quality limits their recognition to the Trefor Rocks natural and quarried exposures. These bioherms are best developed within Morton's Coral Bed as exposed at Eglwyseg Plantation (G. R. SJ.227240) within the basal transgressive MA.2 phase of the Sandy Passage Formation as herein. Two examples occur within MA.3 lithologies; i.e. immediately underlying 'Tf1a' (G. R. SJ 23154331), and immedieately beneath 'Tf.2' (G. R. SJ.23194329).

These bioherms all contain large cerioid or plocoid colonies central to the bioherm 'framework', of either Lithostrotion of decipiens, Lonsdaleia floriformis or Palaeosmilia regia.

The two bioherms below 'Tf.2' (fig. 63) are about 1m diameter and 1m tall comprising an aggregation of colonies within sediments of MA.2 microfacies. It is presumed that coral planulae could readily attach and grow from the hard substrate of another (?dead) coral colony.

From observations of 17 apparent encrusting relationships, Lithostrotion decipiens provided a substrate for 9 encrustations, presumably representing a 'pioneer' species. Encrusting forms include Palaeosmilia regia; Lithostrotion irregulare; L. decipiens; Syringopora spp, Cladochonus sp Aulopora sp; Lonsdaleia duplicata, and chaetetid sponges.

Fig 63 c shows a minor MA.3 bioherm within MA.3 sediments (lithofacies L.1) noteably with lateral sediment-encrusting extensions of Syringopora of ramulosa occurring towards the top of the bioherm. This indicates that sedimentation kept pace with bioherm growth. It is probable that at most one or two colonies were growing at any one time, and there may have been prolonged periods without growth.

The presence of a proportion of overturned coral colonies within and adjacent to these bioherms is significant. All the recorded multispecific bioherms occur within laterally persistent biostromal units, themselves containing a large proportion of overturned colonies (50 - 20% in some biostromes). Within Morton's coral bed the biohermal accumulations apparently occur every 50 - 100m along outcrop in the Eglwyseg Plantation area, visible primarily as large (1m diameter) colonies of Lithostrotion decipiens. At the best exposed of these bioherms (cleared by the author at G. R. SJ.22752404, fig. 64 b) at least 10 colonies (out of 25) are not in growth orientation.

Fig. 63 a shows a transition stage between biostrome and bioherm also from Morton's Coral Bed, in which the coral colonies are densely packed compared to the surrounding sediment but are still sediment supported.

Interpretation.

Larger colonies, especially if embedded within the sediment, would be most resistent to bottom current disturbance. It is envisaged that such colonies provided a baffle, for other locally transported colonies.

With initiation of the colony aggregation, an inertial effect may have both kept colonies being trapped around the bioherm flanks,

(Figs. 63 - 66)

KEY TO NUMBERED CORAL COLONIES

1 Palaeosmilia regia

2 Lithostrotion pauciradiale & L. irregulare

3 Corwenia rugosa

4 Lonsdaleia duplicata

5 Nemistium edmondsi

6 Lithostrotion decipiens & L. maccoyanum

7 Diphyphyllum sp.

8 Lonsdaleia florriformis florriformis (small form)

9 Ld. florriformis subsp.

11 L. junceum

12 L. cf martini

13 Syringopora catenata

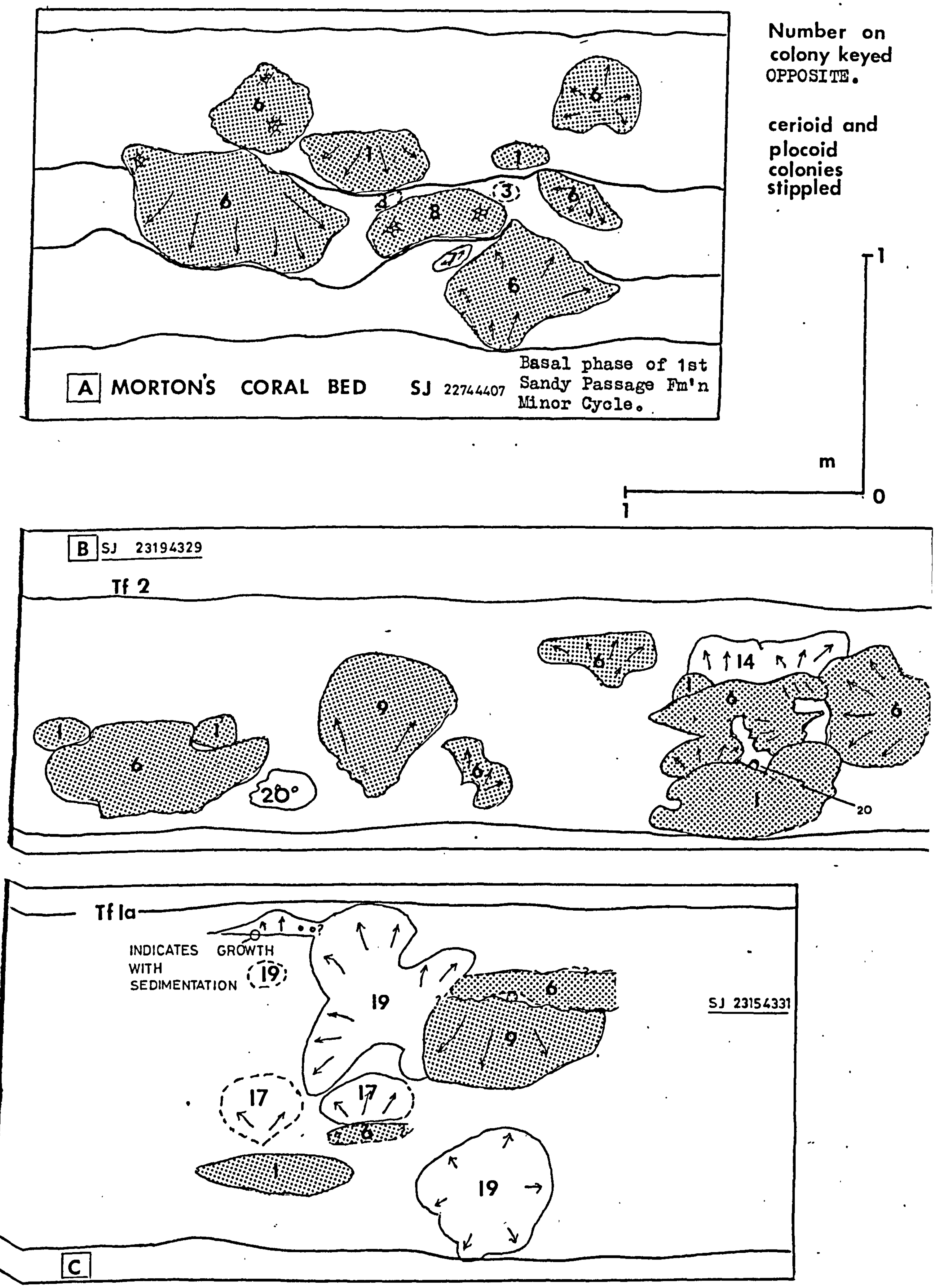
14 S. cf ramulosa

15 S. reticulata & S. cf distans.

These figures are a selection of vertical quadrats of Brigantian coral beds.

MINOR INCIPIENT BIOHERMS (BRIGANTIAN)

Fig 63



GROWTH - BAFFLE CORAL BUILDUPS -- MORTON'S CORAL BED (SANDY PASSAGE FM'N)

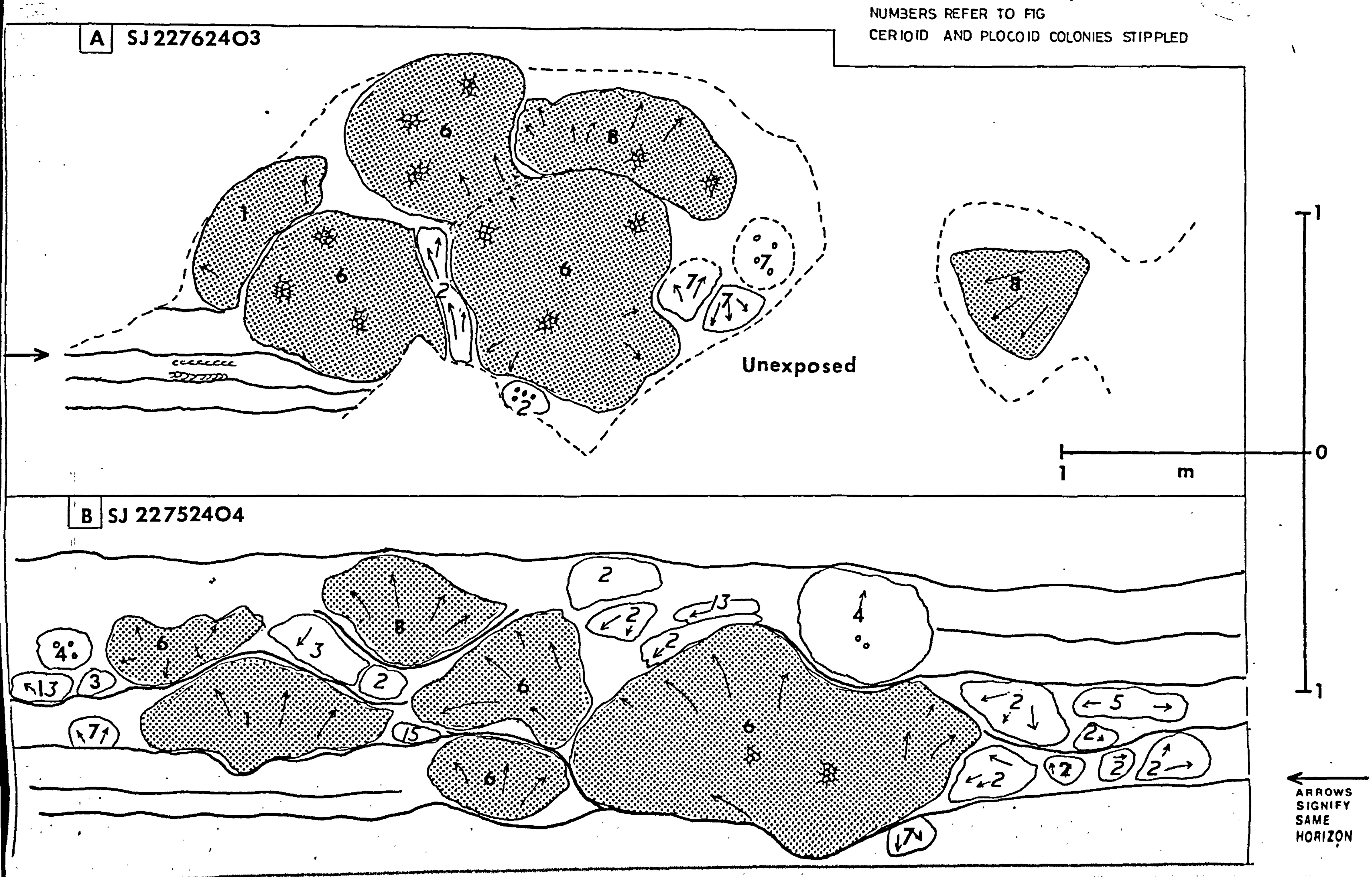
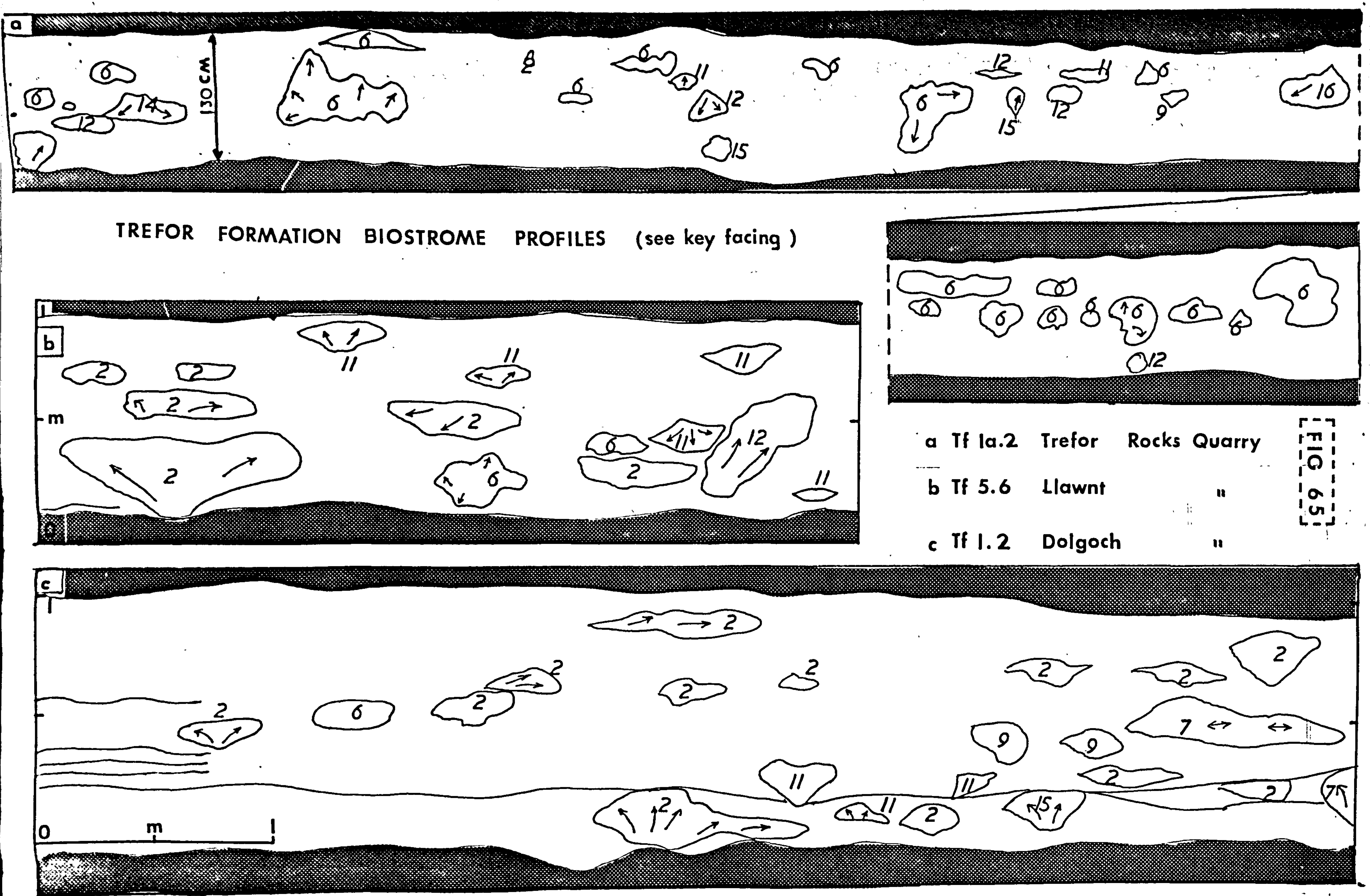


Fig 64



TREFOR FORMATION BIOSTROME PROFILES (see key facing)

a Tf Ia.2 Trefor Rocks Quarry

b Tf 5.6 Llawnt "

c Tf 1.2 Dolgoch "

FIG 6.5

Sketch vertical quadrat of a biostromal level in Pant Hîr Quarry (SJ *23772787*) (L. Brigantian)

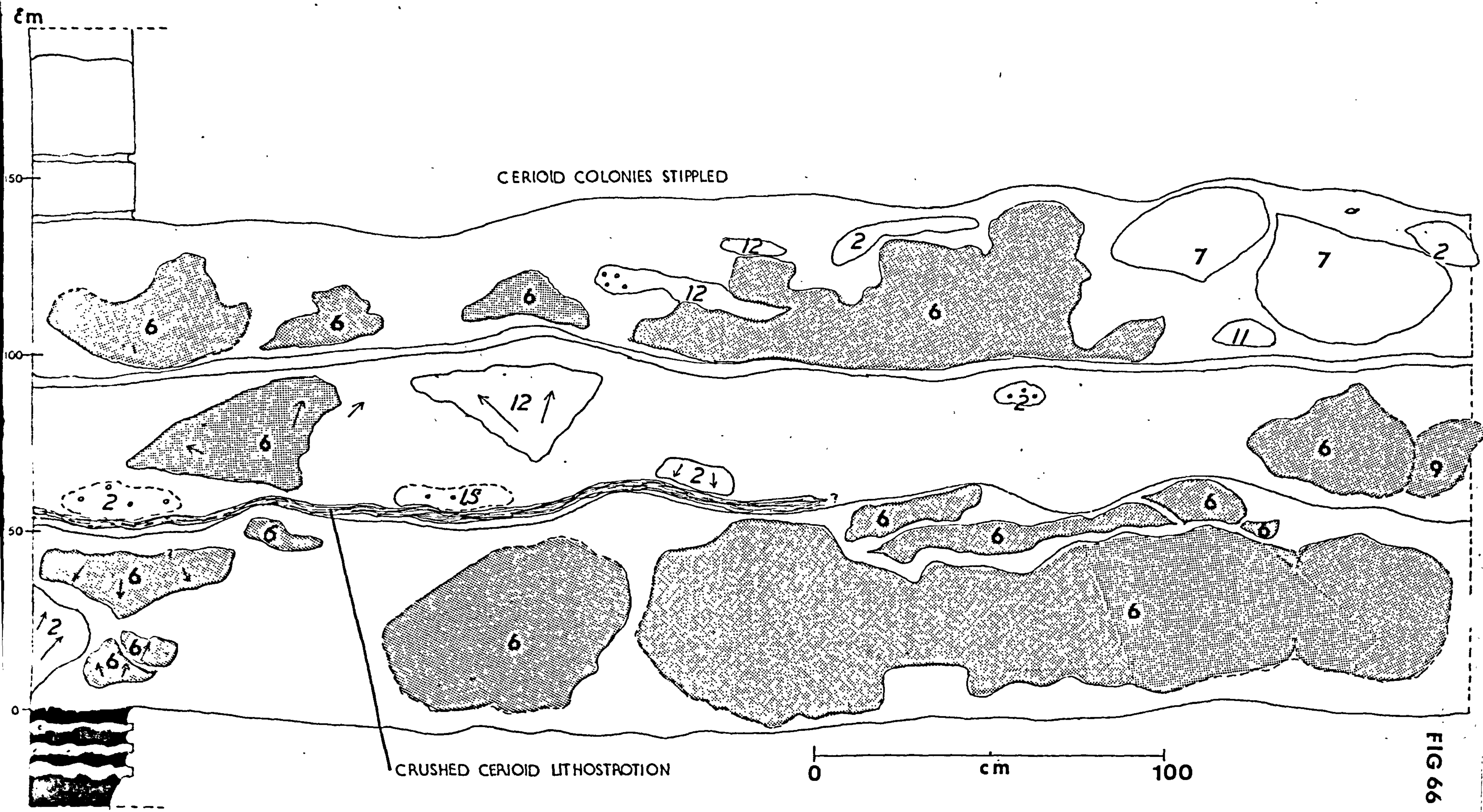


FIG 66

and provided a relatively sheltered habitat for coral planulae to develop upon. Therefore, these 'bioherms' are a product of both in-situ growth and 'baffle' accretion. The Morton's Coral Bed bioherms (Fig. 64) are prime examples of this.

10.4.3. Coral Biostromes.

The best exposed biostrome in the Trefor Formation is between 'Tf.1a' and 'Tf2 on Trefor Rocks, in an 80cm to 200cm thick Lithofacies L.3, argillaceous packstone (MA.2) phase. Measurements from lateral quadrats on vertical faces within this biostrome are given in table 7. Even in this relatively coraliferous unit, however, the vertical face of the biostrome contains <6% by area coral colonies. The cerioid Lithostrotion decipiens dominates the coral assemblage, accounting for almost half the number of colonies.

Figs 65 & 66 indicate the scattered distribution patterns of the coral faunas from other biostromal levels within the Tynant Formation. These biostromes are primarily associated with Lithofacies L.3 units towards minor cycle bases. Noteworthy is the Pant Hir example (fig.66) that can only be traced for 10m into relatively unfossiliferous sediment.

The basal phase of 'Tf.4.5' along Trefor Rocks contains a coral biostrome assemblage with >70% Lithostrotion junceum. L martini and Lonsdaleia floriformis are other minor components. This biostrome occurs along a bedding plane (?indicating a period of non deposition) in which 85% of the colonies are spaced <60cm apart (19 readings)

Many larger in growth cerioid colonies have irregular and indented surface morphologies reflecting growth keeping pace with deposition of the sediment. Within some cerioid colonies sediment patches attest to inactive growth. Sedimentation itself was an important contributor to polyp demise.

Morton's Coral Bed is represented above Minera by a scarp (e.g. G. R. SJ 25275148) with poor exposure, but noteably lacking the dominance of Lithostrotion decipiens. At this locality nearly 50% of visible colonies (about 20 in toto) are Lonsdaleia duplicata, whilst at Trefor Ld duplicata is a minor component. This may reflect local variations in the coral faunas. Within this Minera biostrome large colonies of Ld duplicata up to 1m long and 1m maximum girth, occur as tall cones, many toppled on their sides. Fig. 67 illustrates their dominant orientation (8 readings) with apex pointing NNW, suggesting dislodging bottom currents from this direction. Such currents would have been palaeocoast parallel.

Storm currents recognised as important sediment transporting agents are primarily alongshore to offshore (Walker 1979, p.79). Evidence of Recent storm bed sedimentation is readily destroyed by bioturbation, and this mechanism may have removed evidence of storm beds in Trefor Formation L.3 lithofacies, primarily due to the Zoophycos animal.

The lack of high energy bottom sediments (e.g. rudstone laminae or lenses) and gutter casts suggests that colony transport and disturbance was local. Other plausible methods of colony inversion include bioturbation and persistent scour around and beneath colony margins (cf. Hubbard 1970 and section 8.3.1.1). Scour may have caused colony instability during higher energy bottom current events.

fig 67

Directions of coral colony 'toppling'
?Mortons Coral Bed, Minera

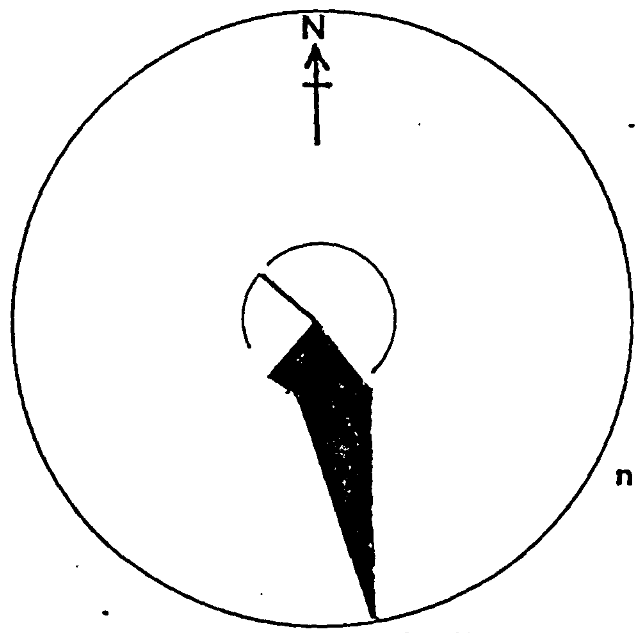


Table 8

SEMI-QUANTITATIVE BRACHIOPOD ORIENTATION DISTRIBUTIONS (BRIGANTIAN)

Dominant Brachiopod	TRANSGRESSIVE PHASE Lithofacies L.2; L.3; L.6				REGRESSIVE PHASE Lithofacies L.1; L. 5.			
	G(%)	O(%)	M(%)	P	G(%)	O(%)	M(%)	P
<u>Semiplanus</u>	100	0	0	10	20	50	30	6
<u>Gigantella/ Gigantoproductus</u>	60	5	35	17	45	45	10	11
<u>Productus</u> sp	75	0	25	4	0	50	50	2

G = individuals dominantly in growth position.

O = individuals dominantly overturned and/or disarti ulated

M = uncertainty, and mixed assemblage of G and O.

P = sample population, total population = 50 shell beds

10.4.4. Brachiopod 'Biostromes' of the Trefor Formation.

The density and diversity of brachiopod associations within the Trefor Formation is very variable. Their diversity is readily appreciable from the appended faunal lists.

In growth orientation shell beds of smaller brachiopods are mostly restricted to argillaceous MA.2(and L.6 sediments). Table 8 indicates semi-quantitative field observations on dominant shell orientations within 50 shell beds scattered over the Brigantian outcrop investigated. Only broad trends are significant due to the paucity and semi-quantitative form of the information. It does, however, reveal trends anticipated from sedimentological evidence alone, ie. the larger robust shells are the most stable MA.3/ MA.1 microfacies components, and MA.2 microfacies contain a higher percentage of shell beds dominated by in-growth position individuals.

Where associated with the massive-bedded MA.3 phases valves tend to be more disarticulated, inverted and less densely packed, reflecting the higher degree of sediment movement and therefore bottom currents in MA.3 environments. (?) (section 9. 3. 1).

10.5. ASPECTS OF EARLY DIAGENESIS OF TREFOR FORMATION SEDIMENTS.

10.5.1. Diagenetic Mottling.

Description. Mottles (of darker sediment within light) up to 4cm diameter, as in many Asbian examples overlie the transition from underlying thin bedded argillaceous sediments (Lithofacies L.3) at the base of massive bedded paler 'regressive' phases (Lithofacies L.1).

One particular mottle horizon (see chart D) within the 'Tf. 2.3.' minor cycle of the Trefor Formation may be readily correlated from Trefor Rocks to Treflach ($\sim$ 20km)

Petrographically the mottles comprise typical MA.3.1, containing up to 50% <u>Coelosporella</u> (e.g. s/m 1360 Llawnt). However, the surrounding, paler intermottle matrix is virtually devoid of <u>Coelosporella</u>, and has much microstylolitisation and compaction, giving a packstone texture and an overall appearance of an intraclastic conglomerate, (Plate 18, fig. F). <u>Coelosporella</u> occurs within the intermottle matrix in pressure shadows on mottle margins, and in areas of less compaction. Specimen 070 (Trefor Rocks, 'Tf.2.3') has a well compacted and microstylolitic intermottle matrix with occasional irregular 50 - 250 µm clear spar patches that contain sub-rounded 'peloids', but lack the shape of <u>Coelosporella</u> thalli (Plate 18 fig D).

The characteristic preservation of Coelosporella within non-mottled sediment is as clear spar, suggestive of void recrystallisation as envisaged by Bathurst (1966) for aragonitic molluscan clasts. Within the mottles, clasts are irregularly orientated, whilst in the intermottle matrix clasts are subparallel to the bedding as a result of compaction.

<u>Interpretation.</u> - The lack of mottles of different lithologies (especially considering their size and lateral extent, suggests they are not intraclasts, but <u>in situ</u> diagenetic and/or biomechanical artifacts. Potential intraclast sources from underlying sediments are, in contrast, often sediments of MA.2 microfacies. The mottle textures resemble the cement rich mottles of Asbian pseudobreccias (see section 9. 2.2.1).

The compositional distinction between mottle and matrix is significant, reflecting either differential diagenetic processes or a primary compositional lamination in the sediment. There is no evidence of such distinct compositional (non-mottle) lamination within the massive MA.3 dominated regressive phase of cycles, and their lateral extent also suggests this is unlikely

Aragonite is a metastable mineralogy within carbonate diagenetic environments, compared to low magnesian calcite. Extant dasyclads secrete aragonite. Coelosporella is a 'typical' dasyclad, with well developed utricles (Wray, 1977). Rare Coelosporella thalli are now ?neomorphic calcite (e.g. s/m 1332, Pant Hir) with an irregular brown dusty calcite mozaic (indicative of organic remnants and thin solution film transformation, cf Hudson, (1962) and Wardlaw et al (1978)). Mottle-associated Coelosporella are clear void filling spar indicative of void-stage transformation.

I suggest that Coelosporella pervaded the primary sediment and underwent void dissolution from an original aragonitic mineralogy, but filling of Coelosporella moulds by clear calcite spar did not occur within intermottle sediment. This mould recementation did, however, occur within the mottles and was probably associated with local sediment cementation defining the extent of the mottles. Subsequent compaction during burial provides a mechanism for destruction of uncemented intermottle Coelosporella moulds.

In Land's (1967) cementation history of Bermudan limestones, aragonite dissolves following early cementation and loss of Mg^{2+}, and is concommitant with reprecipitation as calcite spar.

As Coelosporella thalli are mostly not micritised, their moulds may have been preserved only by adjacent clasts or matrix, or ?earlier cement fabrics (now indiscernable). This mechanism further corroborates the differential cementation of Asbian pseudobreccia mottles of section 9.2.2.1. It appears that this differential cementational history is a recurrent phenomenon within Upper Dinantian shelf carbonates.

Significance of Coelosporella Distribution. The recognition of relict patches of clear spar within the compacted sediment attests to locally incomplete compaction and destruction of *Coelosporella* moulds. Utricles of *Coelosporella* are commonly infilled by homogeneous micrite, appearing as peloid-like bodies against the clear spar cement replacement of the thallus cortex. 'It is these peloid-like' forms that attest to the presence of *Coelosporella* as a primary component within the intermottle sediment.

It may be reasonably expected that other clasts of primary aragonitic mineralogy would also be lost from intermottle sediment if compaction resistent moulds are lacking. If intermottle sediment originally contained up to 50% *Coelosporella* clasts, then the degree of compaction including microstylolitisation is very high. This is recognisable in the apparent 'flow' orientation of intermottle clasts around mottle margins (Plate 18, fig. F).

Part D

SUBAERIAL EMERGENCE AND CYCLE MECHANISMS

11. SUBAERIAL EMERGENCE PHENOMENA

The stratigraphic distribution of subaerial emergence surfaces provide very useful lithostratigraphic marker horizons, especially defining regressive cycle boundaries. Their form and lateral variation also provide evidence of palaeoclimate and palaeoenvironment.

11.1. SUTURED DISCONTINUITY SURFACES

11.1.1. Description.

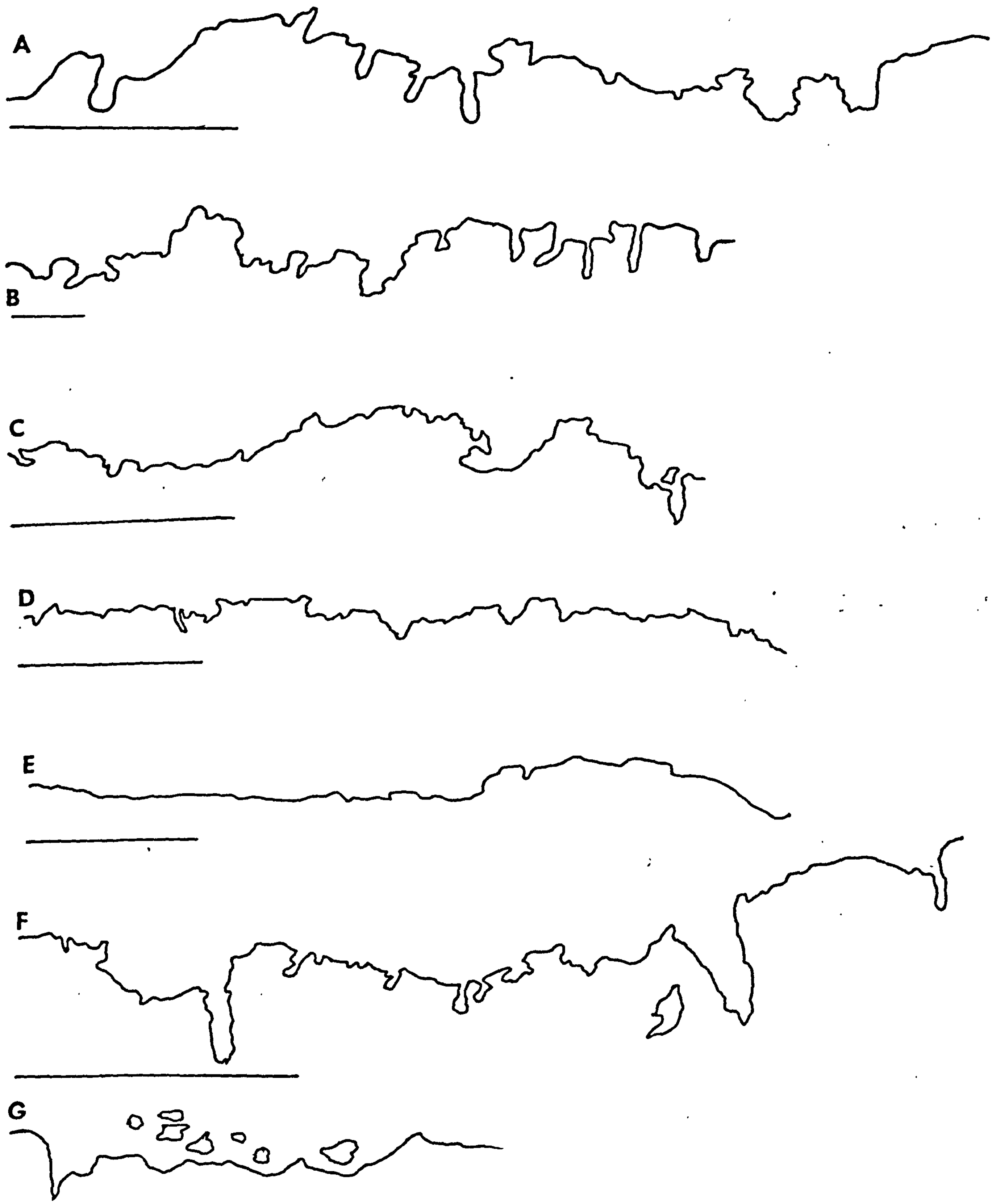
Sutured discontinuity surfaces (S.D. Surfaces) are defined here as irregular non-stylolitic planes that have a 'sutured' (crenulate) appearance in vertical profile, with a relief from 2cm (s/m 2570, Llanymynech Member†) to greater than 20cm (Plate 19 fig.A,B,C fig 68). They occur within both Asbian and Brigantian strata, and usually define abrupt lithological transitions. They mark a period of both non-deposition and modification of the sediment surface, but are readily distinguished from palaeokarstic surfaces by their small scale and angular irregularity, lack of smooth undulations, presence of overhang features, and lack of associated calcrete structures and palaeosols.

Although these surfaces superficially resemble denticular stylolites (Bushinskiy,1961) they lack coatings of insoluble clay residues, and 'slicklolite' (stylolite slickenside) striae.

They always overlie sediments of MA.1 lithology, many with tubular and medium to small irregular fenestrae. Large intraclasts of underlying sediments usually overlie them, mostly in a matrix-supported MA.4.2 manner (section 7. 4. 2.) with MA.2 or MA.3 matrix. They also occur within MA.1 units (e.g. fig.69),

† Plate 14, fig. B

SUTURED DISCONTINUITY SURFACE MORPHOLOGIES



Scale bars 10cm

A–F from field photos

PLATE 19 Subaerial Emergence Surfaces

A Sutured discontinuity (S.D.) surface (outlined). Note paucity of overlying intraclasts. This is interpreted as an intertidal erosion - dissolution surface (and minor cycle boundary) formed through mechanical, biological and chemical processes. Tynant Formation, Tynant, G.R. SJ 21562197. P = prominent high- point

B Close-up of A. Note irregular vertical and horizontal borings.

C Multiple emergence surface: Lower, irregular bored / burrowed (?Skolithos s.l.) surface with slight?sutured discontinuity relief, overlain by a planar erosion surface, here merging laterally into an S.D. surface (to right) with minor relief. Surfaces outlined. Sychtyn Member, Nant Mawr Quarry, G. R. SJ 25202500

D Form 'A' mamillated palaeokarst. 'Eg.5', Eglwyseg Formation, Tynant, G. R. SJ.22194533.

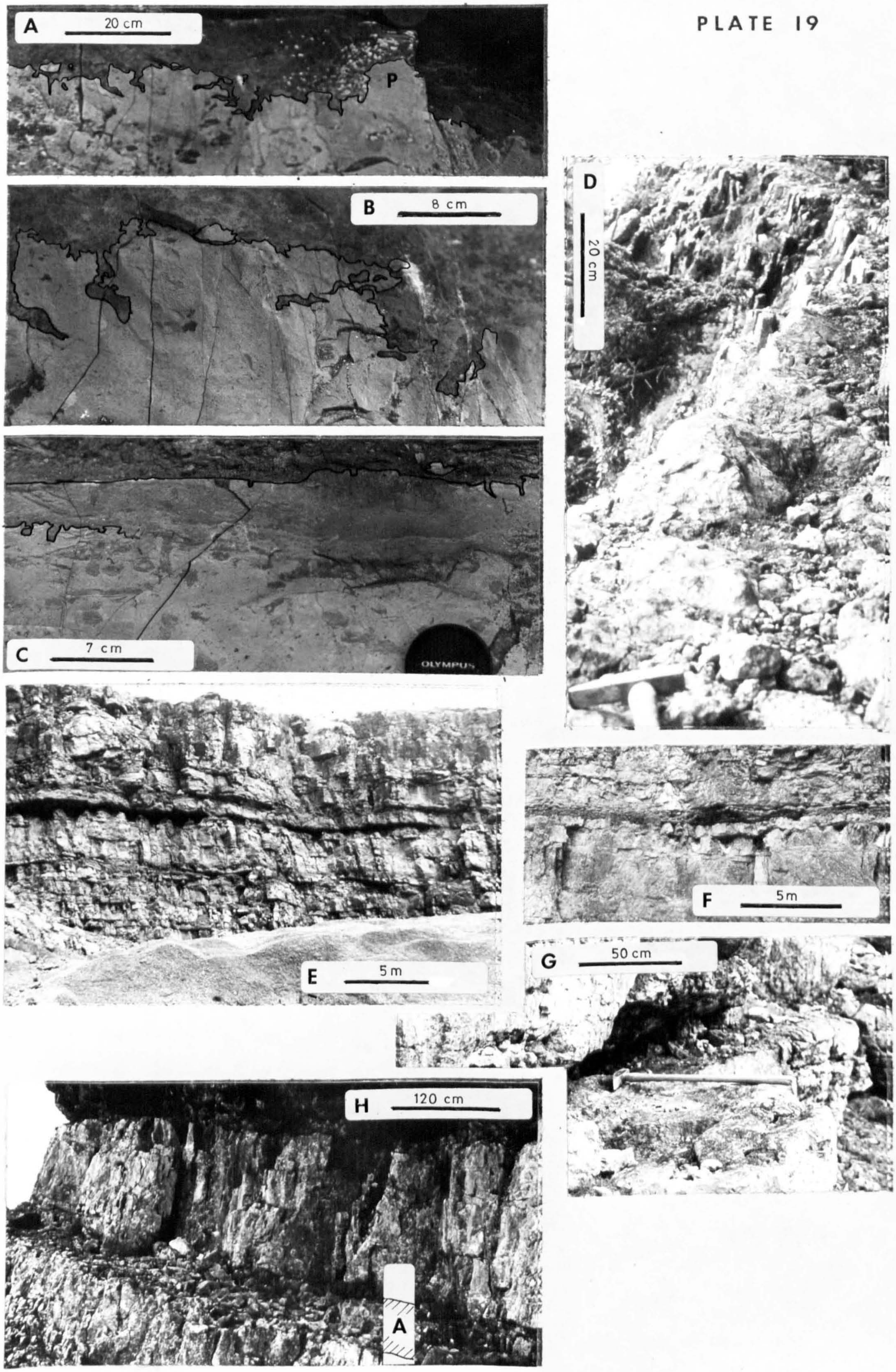
E Form 'B' "potted" palaeokarst, Ly.2, Llynclys Formation, Nant Mawr Quarry, G. R. SJ.25252500

F Same surface as E, 1km SW of Nant Mawr, G.R. SJ26552438

G Form 'A' 'palaeokarst' smooth surface, ~10cm amplitude. Tubular calcrete structures are preserved on 'hummocks'. Eglwyseg Formation, Tan y Graig. Eg.9. G.R. SJ 22454672.

H Form A ?palaeokarst (A) (with underlying calcrete phenomena), Eglwyseg Formation, Eg.4 Tynant, G.R. SJ 22044549.

PLATE 19



there overlain by thin intraformational conglomerates. The volume of conglomeratic intraclasts overlying S.D. surfaces does not account for the degree of relief observed on many of the surfaces, implying that more sediment has been removed than locally deposited. Their angular and irregular surface morphology is accentuated by irregular, subvertical borings. Rarely Chondrites sp. underlies the surface (e.g. s/m 2570). The subvertical borings (Fig. 68.) are infilled with overlying sediment, and sharply truncate sediment lamination, although no individual grain truncations have been observed. back-fill spreiten are visible. Chondrites 'burrows' are also infilled with overlying sediment but they are probably 'push-apart' structures. No encrusting organisms have been observed on any surface.

Both bladed and micritic early cement fringes rarely line fenestral cavities underlying S.D. surfaces (e.g. s/m 029 Tynant Formation). The surfaces extend over metres to tens of metres, and are laterally equivalent to stylolites, planar surfaces† and palaeokarsts (fig. 32). They are most common within the Tynant and Whitehaven Formations, but rarely occur within both the Eglwyseg and Trefor Formations. In the Sychtyn Member they often occur above or below palaeokarsts (fig.69 vi.i,ii). Poor exposure due to their position on lithological boundaries (where recent weathering, vegetation and scree accumulation heavily affect exposed sections) and low amplitude relief hinders their recognition.

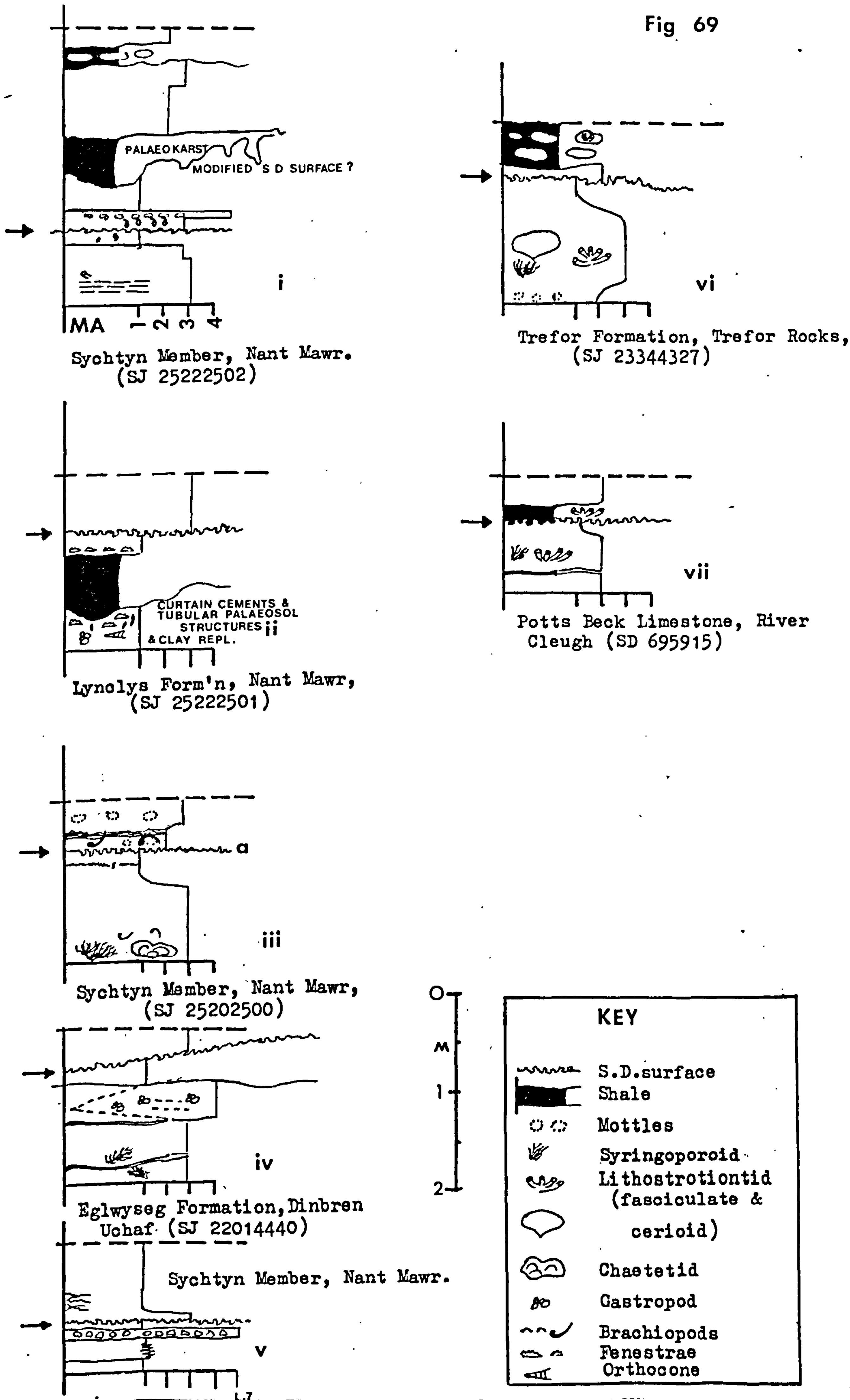
11.1.2. Interpretation of S.D. Surfaces

Bromley (1975) defined the term 'discontinuity surface' as a small scale, chiefly intraformational break in the sedimentary record. Janussen (1961) noted that in carbonate lithofacies these discontinuity surfaces were always present in potentially emergent

† Plate 19, fig. C

SUTURED DISCONTINUITY SURFACE : FIELD RELATIONSHIPS :

Fig 69



positions on shallow marine shelves.

The presence of surface relief, sharp angles and intraformational conglomerates implies that the underlying lime muds were at least partly lithified during their formation. Section 6. 4. 7 describes plausible cementation mechanisms. Intertidal erosion, described by Hardie and Ginsburg (1977) (section 7.4.5.1) may have initiated S.D. Surfaces.

Hardgrounds (sensu Fursich, 1979) are the product of chemical, mechanical and biological processes. Their morphology parallels that exhibited by S.D. surfaces, but whether these have common origins must be clarified.

Although many hardgrounds are formed in grain-supported matrix-free carbonates, their micritic counterparts have been described (e.g. Tucker, 1973, 1974). R. G. C. Bathurst (lecture delivered at the Palaeontological Association Annual Meeting, Reading, 1977) stated that (in clean sand environments) hardground cementation was probably initiated slightly below the sediment-water interface, where a balance between grain-grain stability, pore water saturation and flow is attained. Overlying sediments are commonly of similar lithologies, lacking early cement fabrics, in contrast to the marked lithological change observed over S.D. surfaces.

Hardgrounds are normally associated with an assemblage of encrusting epifauna and boring infauna, especially in shallow carbonate shelves, from the Ordovician to the Recent (e.g. Palmer and Palmer,1977;Purser 1969 Bromley, 1968; Fursich 1979).

Early isopachous cement fringes characterise Recent shoaling environment hardgrounds, however only rarely are early cement fringes preserved in penecontemporary porosity of S.D. surface sediments.

All S. D. surfaces occur upon sediments of MA.1 peritidal suite, many with evidence of penecontemporary emergence from fenestral fabrics (see section 6. 4. 2), and at times laterally equivalent to palaeokarsts.

The erosion (part of the formation process) of S.D. surfaces is, at least in part, associated with the transgressive event, indicated by the overlying conglomeratic sediments. The S.D. surfaces are, therefore, partly analogous to the " rock grounds" of Fursich (1979).

A recent analogue of these S.D. surfaces occurs on intertidal rock platforms of Bikini Atoll (Revelle and Emery, 1957), forming by dissolution of lithified carbonate in irregular basins which are readily infilled with sands. Read and Grover (1977) used these dissolution features as analogues of their scalloped erosion surfaces. Revelle and Emery (1957, p.705) stated that intertidal dissolution was a rapid process, faster than meteoric karstic dissolution, and have shown that it is diurnally controlled by biochemical deoxygenation and increase in CO_2 partial pressures within the isolated intertidal pools.

This model requires pre-lithification of the sediment, although Revelle and Emery (op.cit) have shown that even where active diurnal dissolution is occurring, the intertidal ponds may be supersaturated with respect to $CaCO_3$.

Slight erosive scour occurs in these intertidal pools by movement of entrapped sand. Such scouring may modify surfaces extensively, especially during an active transgressive event. Storm erosion may be responsible for the intraformational conglomerates associated with these Carboniferous S.D. surfaces.

S/m 2528 (upper regressive phase of Tynant Fmn, fig. 9 , section 6.4.1.5) displays pervasive fenestral scalloped dissolution features. (fig. 42). This may be a further example of intertidal dissolution.

Read and Grover's (1977) scalloped erosion surfaces occur on top of fenestral limestone units, are overlain by 'open marine skeletal limestone units' and pass laterally over a few metres into planar erosion surfaces also.

Although this setting is very similar to Read & Grover's scalloped surfaces, the S. D. surfaces have morphological dissimilarities. S. D. surfaces lack a smooth, concave surface expression with sharp intervening pinacles. In a few examples, however (e.g. fig.68) undulose relief on a metre wavelength occurs.

S. D. surfaces were most probably formed post early lithification by a combination of tidal dissolution, mechanical erosion, and biogenic activity. Ramsbottom (1974 p.59) described 'hardgrounds' from the Potts Beck Limestone (section 5. 1. 1), here interpreted as S. D. surfaces. No Lower Carboniferous counterparts of the encrusted and modified hardgrounds sensu Fursich (1979) are known to the author.

7.1.1.3. Palaeogeographic significance of S. D. surfaces.

S. D. surfaces, formed in intertidal environments, Theoretically this environment is laterally equivalent to both subaerially emerged supratidal and subtidal sediments. Their geographic distribution should indicate whether they are transgressive or regressive phenomena. If they were formed on transgressive events, their distribution along cycle boundaries would be persistent and widespread as the transgressive shoreline would modify pre-existing supratidal platforms with associated palaeokarstic surfaces. If they were formed during regression, they would be modified by later supratidal karstification, and subsequent transgressions. Their distribution would, therefore, be limited (laterally) to a narrow zone, paralleling the strand line and delineating the position of maximum regression. This is shown diagramatically in fig. 70 .

The evidence from their distribution suggests the latter situation. They do not have a persistent lateral extent. Rarely they provide ?correlations over < 2km of outcrop (subparallel to the depositional strike) and mostly are limited to hundreds of metres or less. Where preserved, erosion on the succeeding transgression has partly

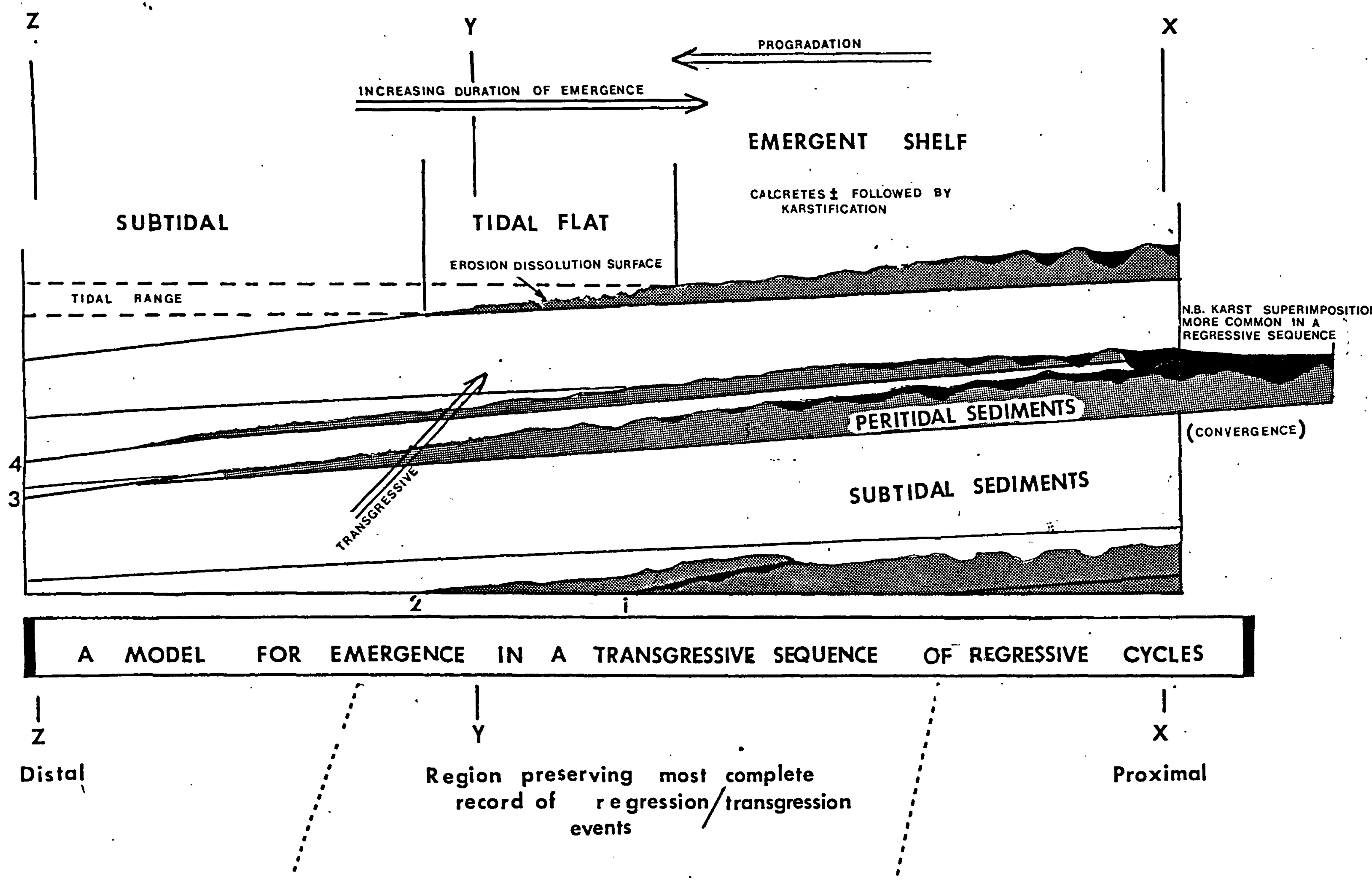

Z
Y
X
PROGRADATION
INCREASING DURATION OF EMERGENCE
EMERGENT SHELF
CALCRETES ± FOLLOWED BY KARSTIFICATION
SUBTIDAL
TIDAL FLAT
EROSION DISSOLUTION SURFACE
TIDAL RANGE
N.B. KARST SUPERIMPOSITION MORE COMMON IN A REGRESSIVE SEQUENCE
PERITIDAL SEDIMENTS
(CONVERGENCE)
4
3
TRANSGRESSIVE
SUBTIDAL SEDIMENTS
2
1
A MODEL FOR EMERGENCE IN A TRANSGRESSIVE SEQUENCE OF REGRESSIVE CYCLES
Z
Distal
Y
Region preserving most complete record of regression/transgression events
X
Proximal

fig 70

modified them. In the Sychtyn Member, especially, palaeokarsts are vertically associated (see Chart C.) whilst in the Trefor Formation (fig. 32) one S. D. surface can be traced from a position central to the Llangollen Embayment, southwards into a palaeokarst, and planar surface with calcrete phenomena, indicating the maximum extent and development of a cycle regression.(Also see Mesothem D5a/D5b, fig.49)

11.2. PALAEOKARSTIC SURFACES.

Somerville (1977, 1979a, 1979C) recognised palaeokarstic surfaces as conspicuous irregular upper surfaces to limestone beds,at times infilled by a palaeosol, positioned on regressive cycle boundaries of the Eglwyseg and Trefor Formations.

These surfaces are valuable correlation aids (e.g. fig 22) and the variability of both their relief and distribution provide information on shelf palaeogeography, timing and duration of subaerial emergence.

The four forms of palaeokarstic surface, described below, are intergradational with each other, but this classification appreciates the variation encountered both in the surface morphologies, and the factors controlling their formation.

11.2.1. Surface Form (Classification summarised in fig. 71)

11.2.1.1. IRREGULAR MAMMILATED KARST (FORM A.) These surfaces are undulose bedforms with an irregular vertical profile of close-spaced, low hummocks or ridges and low intervening flat bottomed or saucer shaped depressions, with 10 to 50 cm relief (Plate 19 fig. D). They are the most common palaeokarst type in the North Wales succession, and many do not have clay infillings. Large depressions, up to 50cm deep and 10m wide may 'ornament' the surface (e.g. Tynant/Eglwyseg Formation junction, Minera Quarry, Chart E.).

Sediments underlying the palaeokarstic surfaces are only stained (principally maroon) where maroon and green palaeosols are present.

MORPHOLOGICAL CLASSIFICATION OF PALAEOKARSTS USED HEREIN

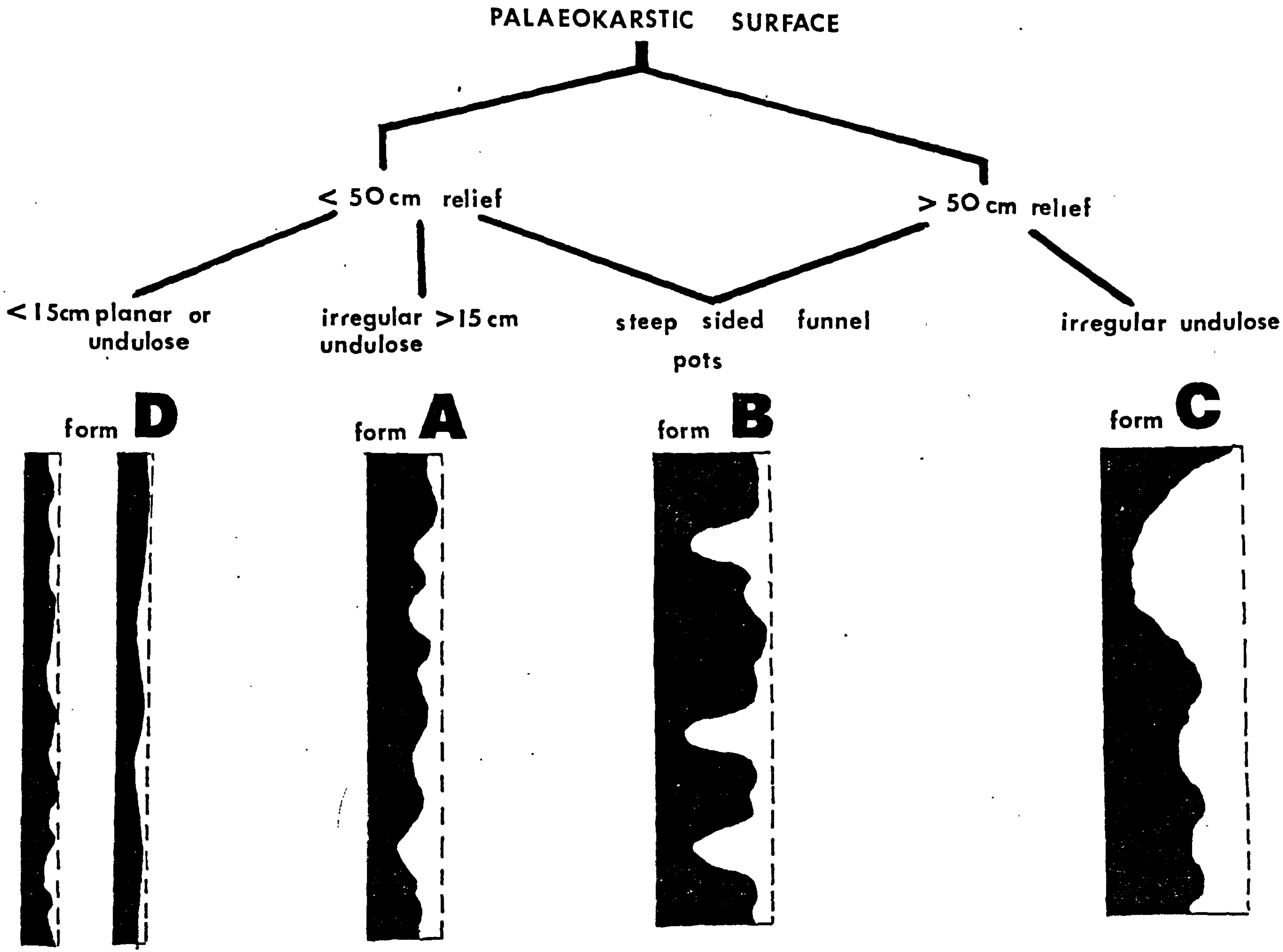


Fig 71

Unlike Read and Grover's (1977) scalloped erosion surfaces, they lack sharp pinnacles and concave-up 'scalloped' appearance. Overlying carbonate beds thicken into the depressions where clay infills are absent. Both laminated calcrete crusts and tubular calcrete structures may occur within the underlying sediment, normally concentrated towards hummock tops, but within both Asbian and Brigantian strata, they are often absent. These surfaces can occasionally be traced for > 5km (e.g. Eg. 5, Eglwyseg Formation, fig. 22) without change in palaeokarst style. Meal Bank Quarry, Ingleton, has well exhumed surfaces of this form (Kingsdale Limestone, Asbian, see Chart B.). Form A is equivalent to the "gently rolling palaeokarstic surface, type 1" of Walkden, (1974.)

11.2.1.2. <u>REGULARLY 'POTTED' KARST (FORM B)</u>. Only one example of this surface type has been discovered, in the Llynclys Form'n, Nant Mawr, (G. R. SJ. 25252500 Plate 19 fig.E& F). The surface is very distinctive with deep narrow <u>funnel shaped 'pots'</u>, up to 100cm deep and 50 - 100cm diameter, with rounded bases and nearly vertical sides, separated by† a flat or slightly undulose form A palaeokarst. The pots in this particular Sychtyn Member example range from 1m to 10m apart (in vertical section) and are infilled with grey and green clays (?palaeosols). This surface resembles Walkden's (1974) 'type 2' palaeokarst.

11.2.1.3. <u>DEEP IRREGULAR KARST (FORM C)</u>. Examples of this form occur within the Eglwyseg Formation exposed between Tan y Castell and Bron Heulog, and within isolated positions on karst 'Form A'. Form C has <u>deep karstic relief</u> (> 50cm) (up to 5m on boundary Eg.6 Bron Heulog/ Trefor Rocks, (Plate 2, fig. C)), and none have been observed with calcrete phenomena. The relief has hummocky undulation (Form A) upon it, with depressions from 2m to 20+m wide, which, in the Eglwyseg <u>Formation, is infilled by maroon mudstones</u> and conglomerates (e.g.fig.24a).

† and merging laterally into Form A palaeokarst

Staining of underlying limestone is rare, and limited to reddening around stylolite planes and stylonodules.

11.2.1.4. PLANAR 'KARST' (FORM D). Within the Eglwyseg Formation, at many sites, smooth, but prominent 'master bedding planes' occur, underlain by calcrete phenomena, but lacking a typical karstic profile. Their surface relief is generally less than 10cm, although this may be superimposed on large wavelength ($\gtrsim$ 1m) low relief undulations e.g. boundary Eg9 Tan y Graig (Plate 19 fig. A &G). The calcrete crusts and tubular infestations are developed in pockets, probably caused by erosional scour of an original karst, with the calcretes preserved beneath original depression bases.

Many planar surfaces occur on cycle boundaries in the Tynant and Whitehaven Formations. Some can readily be traced laterally (Sychtyn Member) into undulose karst forms (Form A) or S. D. surfaces, but the majority cannot. In the Llanymynech Member these surfaces are mostly overlain by a thin maroon dolomitic shale (1 - 5cm thick), occasionally containing a coral-brachiopod assemblage. One of these is underlain by calcretes, but calcretes are very rare throughout both the Tynant and Whitehaven formations.

11.2.2. Recognition of Palaeokarstic Surfaces.

Palaeokarstic surfaces were first recognised (one "potted surface") in the Oswestry succession by Wedd et al. (1929), although the significance was not discussed. Previously potholed and irregular bed surfaces had been recorded in the Carboniferous of Derbyshire (in Walkden 1974, p.1232), but they (e.g. Bemrose, 1907) considered the surfaces as secondary, resulting from dissolution of the limestone beneath decomposing clay or volcanic material. Greenly (1900) described sandstone pipes infilling depressions within underlying limestones from the Anglesey Carboniferous, interpreted their formation during a period of shallowing (and possible exposure of the sea floor), likened them to potholes, and suggested a dissolution mechanism but did not directly implicate karstic processes. Dixon (1909) exemplified a contemporaneous solution unconformity in mid-Avonion strata of Pembrokeshire recognising the subaerial origin of such surfaces.

Sargent (1912) used the term "mammillated surface" to describe surface morphologies (?palaeokarstic) at Crick. Cope (1939) inferred a contemporaneous origin for the Derbyshire surfaces, amplified by subsequent workers. Thomas (1953) interpreted intra-formational 'pipes' in D_1 limestones from Glamorgan as a result of subaerial erosion. Walkden (1970) cited distinctions between palaeokarstic surfaces and stylolite morphologies in the Derbyshire D_1 and D_2 (Asbian and Brigantian), concluding that some stylolitic planes modified original discontinuities with overlying clay 'wayboards', to which stylolitic insoluble residues were added. Palaeokarsts in the Derbyshire Upper Dinantian are restricted to shelf limestones, and absent from those of more basinal facies, Walkden (1974, p.1235 - 6). Baughen and Walsh (1980) suggested that sandstone-pipes in the Upper Diantian of Anglesey were formed by palaeokarstic solution beneath a periodically emergent quartz sand cover, with concommitant infilling by this overlying sand. This mechanism implies that the underlying limestone need not have been subaerially emergent following its deposition.

Somerville (1977) noted that beds overlying palaeokarsts were flat bottomed, due to the infilling of karstic relief by clay palaeosols, and cited the development of karstic relief independant of later jointing and faulting and always normal to the bedding planes, even in inclined strata, as evidence of their syngenetic origin. Walkden (1974, p. 1237) described erosive truncation of fossils within underlying sediments, indicating that the surfaces are erosive and not primary.

It is the nature of the bedrock beneath the palaeokarstic surface and the relief infill that indicates whether the surfaces were subaerially emergent, or as proposed by Baughan and Walsh (1980) formed beneath a sediment cover.

Arguments against this latter proposal for the Asbian and Brigantian palaeokarsts are:

1. The 'pot' infillings are often palaeosols or intraformational conglomerates that occur within the confines of the 'pots' themselves, and not contiguous with overlying lithologies.
2. The bedrock often shows evidence of contemporary exposure with calcrete phenomena developed upon and within it.
3. 'Pot infillings contain primary sediment lamination that wedges towards the sides of the pot, draping and wedging against the surface contours. Occasionally, depressions are filled by laterally impersistent carbonate beds.

11.2.3. <u>The Karst Forming Processes.</u>

Karstic surfaces form through solution by acidic waters, normally of meteoric derivation (pH < 7.8 according to Krumbein & Garells, 1952). Recent karstic dissolution of Bermudan Pleistocene carbonates occurs on surficial partly lithified crusts. The origin of this carbonate cement may be from meteoric leaching and dissolution of aragonite within the surficial carbonate (Bathurst, 1975, p.330). Such cementation is greatly controlled by the position of water tables. About 20% porosity is still retained until later introduction of carbonate by e.g. pressure solution. Typical MA. 3 microfacies have between 20 and 40% readily visible cement. Although evidence of early cements (especially vadose) is limited to a very few karstic profiles (section 6. 4. 11) low Mg-calcite void filling cement may not retain any petrographically recognisable "early cement" characteristics. Cathodoluminescence and isotopic studies may improve our understanding of this phase of cementation.

Karst formation usually occurs beneath a thin soil cover, in part derived from insoluble residues of the limestone, and also, as shown by Walkden (1972) and Somerville (1977) from wind-derived volcanic dust. Bonte (1963) believed that soil cover is essential for karst formation. As many palaeokarsts in the Eglwyseg Formation lack palaeosols, Somerville (1977) evoked erosional deflation prior to transgressive sedimentation.

Dissolution of limestone in spray zone and supra-littoral environments of Puerto Rico (Kaye, 1959) causes surface relief dissimilar to the karsts described here. These coastal situations, without soil covers, produce 'jagged' karst profiles, with smooth-round or flat bottomed pits, or deep funnels (coastal lapiés of Tricart, 1972), separated by irregular, sharp ridges. Kaye (1959, p. 85) advocated meteoric waters as solution agents for such karst morphologies which are similar to the 'scalloped erosion surfaces' of Read and Grover (1977). Covered karsts tend to produce more rounded profiles, on both microscopic and macroscopic scales.

Walkden (1974) suggested that high water tables may have accounted for the low relief and high variety of Carboniferous palaeokarsts (due to either a perched water table or, more probably in the context of these surfaces, to their proximity to sea level).

Walkden (op. cit) used data available from surveys of Recent tropical/subtropical karst environments and Pleistocene analogues to estimate the length of time Derbyshire Asbian and Brigantian palaeokarsts took to form. His estimates were between 30 000 and 100 000 years for a surface with average overall reduction of 50cm. The problem is enhanced by lack of knowledge of both rainfall and temperature distribution of the Lower Carboniferous.

The absolute minimum length of time a karst profile takes to develop can be shown theoretically:

Fresh rainwater, with an atmospheric partial pressure of CO_2 (3.5×10^{-4} bars) can dissolve 0.044g of $CaCO_3$ per litre at 25°C (Miller, 1952, p.195) (confirmed by Krauskopf (1967) who stated that in aqueous solution carrying CO_2 in near-surface porosity, $CaCO_3$ content is unlikely to exceed 0.05g/l at 25°C)

Carbonate sands, depending on the grain shapes and sizes have a primary porosity up to 80% by volume; 50% is common. In the bioclastic and peloidal grainstones (MA.3 microfacies) characteristically underlying palaeokarsts within the Eglwyseg Formation, initial porosities may have been as high as 50%, although they are now probably < 2%. Land (1967) described five stages in subaerial cementation of Bermudan Pleistocene limestones from initially unconsolidated sediment. The final stage was recognised as a rock composed totally of calcite with about 20% porosity, after all metastable mineralogies (aragonite and high magnesian calcite) had undergone transformation.

At a first approximation, assuming that the Eglwyseg Formation MA.3 sediments had a porosity of about 20% during (or towards the end of) karstification, then the following calculation holds:

The density of calcite sediment with 20% porosity is $2.15 g/cm^3$ However, organic fractions and submicroscopic (impermeable) pore-space will reduce individual grain density to slightly less than $2.7 g/cm^3$:- Using an initial sediment density of $2.0 g/cm^3$:

1000cm of rainfall (pCO_2 3.5×10^{-4} bars) will dissolve .022cm of 20% porous sediment at 25°C given optimum conditions to allow all the meteoric waters to reach saturation.

i.e. 45 500cm of rain will dissolve 1cm of 20% porosity limestone in ideal conditions.

QUALIFYING CRITERIA:

1. Porespace during karstification decreases as recementation occurs at lower levels and within the phreatic zone.
2. Rapid removal of surface waters will not allow attainment of $CaCO_3$ saturation, although an overlying soil may retain waters and allow more gradual downward seepage.
3. This model does not take into account evaporative removal of surface waters that in most tropical or sub tropical environments is significant and in semi-arid climates will readily exceed rainfall. This model is most suitable for temperate climates (Engh 1980)

These three factors suggest rates of karstic solution much slower than estimated from the above data, but biogenic weathering (increasing pCO_2 & humic acids in soils (Sweeting 1966), high humidities and lower temperatures would go part way to counteracting the above factors.

Possibly comparable climatic situations to the Asbian of North Wales (situated at a palaeolatitude of $\sim 10^0$ south according to Smith et.al., 1973) range from the Caribbean (Andros) with an average annual rainfall of 114cm/year (Ginsburg and Hardie, 1975) to Shark Bay's 23cm/year (Hagan and Logan, 1975) . (Sabkha environments of Abu Dhabi, not readily comparable to the North Wales Asbian have less than 4cm/yr, with annual evaporation rates in excess of 900cm/year (Schneider, 1975).)

Comparing known karstic dissolution rates, cited in Walkden (1974, p. 1244) the theoretical ones calculated above are much more rapid (as predicted) than the $16mm/10^3$ years for Yucatan karsts to $5mm/10^3$ years in Florida (Corbel 1959). Calculations based on the yearly average rainfall of 120cm in both these areas indicate maximum karstic removal of 21mm/10^3 years from $CaCO_3$ saturation data (assuming a density of 2.5g/cm^3 for the karstified limestones).

Most Eglwyseg Formation karsts (form A) have an average relief of $\geqslant$ 15cm, and therefore as minimum time of formation, between c.6 000 years (with 120cm/year rainfall) and 15 000 years (at 50cm/year rainfall), although comparing rates of recent karsts, they may be up to four times greater (assuming atmospheric chemistries similar to Recent.).

The karstic rates (maximum theoretical and observed/inferred) are plotted on the Engh (1980) solutional erosion curve in fig. 72.

Form C surfaces, with average relief up to 2m may have taken a minimum of between 200 000 years (at 50cm/year rainfall(and 83 000 years (at 120cm/year rainfall) although figures considerably higher are probable. Therefore, the mid-Eglwyseg Formation boundary Eg.6 is locally an important non-depositional hiatus that may be of use in inter-shelf area correlations.

Form D palaeokarsts however, do not readily lend themselves to this analysis, as planation erosion associated with the succeeding transgression appears to have been a modifying effect (see on). S.D. surfaces, by analogy with Recent intertidal surfaces probably took less than 4 000 years to form.

Palaeokarsts overlying MA.1 sediments (e.g. Sychtyn Member) may have had a lower syn-karstification porosity than MA.3 sediments once partly lithified (although Recent aragonite muds have a primary porosity of about 50% (Bathurst 1975, p.504)) giving them a slower rate of dissolution than MA.3 counterparts, but the increased ability of fine-grained mud/micrite to enter solution would counteract this.

11.3. CALCRETE PHENOMENA.

Calcretes are formed by subaerial alteration, accretion and

FIG 72

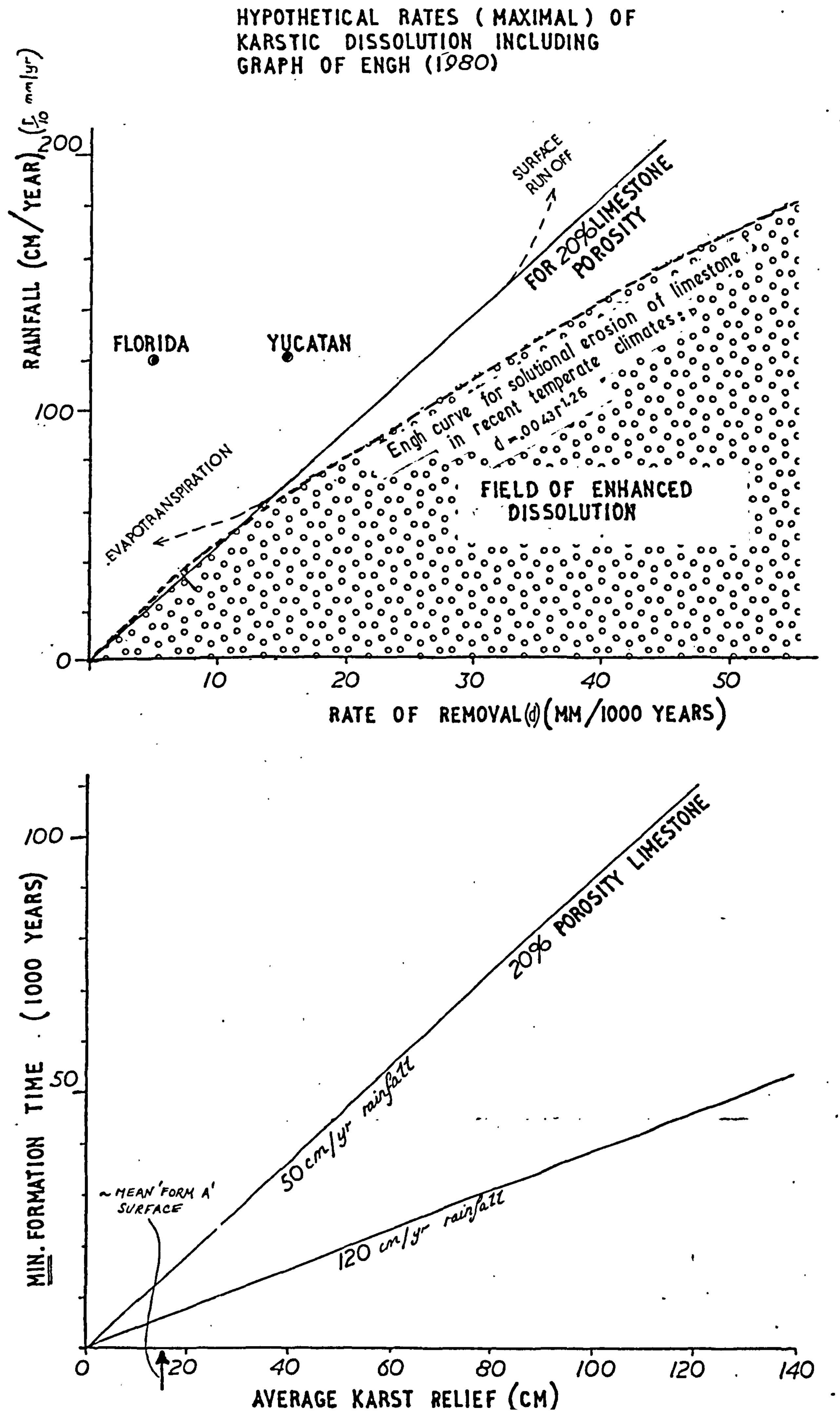

HYPOTHETICAL RATES (MAXIMAL) OF KARSTIC DISSOLUTION INCLUDING GRAPH OF ENGH (1980)
RAINFALL (CM/YEAR)
200
100
0
10
20
30
40
50
RATE OF REMOVAL (d) (MM/1000 YEARS)
FLORIDA
YUCATAN
SURFACE RUN OFF
FOR 20% LIMESTONE POROSITY
Engh curve for solutional erosion of limestone in recent temperate climates: d = .0043r1.26
EVAPOTRANSPIRATION
FIELD OF ENHANCED DISSOLUTION
MIN. FORMATION TIME (1000 YEARS)
100
50
0
20
40
60
80
100
120
140
AVERAGE KARST RELIEF (CM)
20% POROSITY LIMESTONE
50 cm/yr rainfall
120 cm/yr rainfall
~ MEAN 'FORM A' SURFACE

PLATE 20

A Reworked clast of laminated calcrete crust underlying a thin crust with associated tubular calcrete structures s/m. 2184. Immediately beneath 'Eg.2a'. Trefor Rocks,. Note: underlying crust clast from main crust, 20cm beneath. Polished block.

B Autobreccia with clasts coated by dense laminated calcrete, mature profile, Brigantian, Graig Quarry (Chart D)

C Neptunian "fissure" of calcrete underlying a laterally persistent laminated calcrete crust, loose block Bron Heulog Quarry ?from Eg. 8., Eglwyseg Formation.

D Fine micrite laminae within the millimetre laminae of a porous laminated crust, s/m 1313 G. R. SJ 221445

E Porous laminated crust with tubular structures, infilled by alveolar structures, s/m 1313 G. R. SJ221445.

PLATE 20

A Reworked clast of laminated calcrete crust underlying a thin crust with associated tubular calcrete structures s/m. 2184. Immediately beneath 'Eg.2a'. Trefor Rocks,. Note: underlying crust clast from main crust, 20cm beneath. Polished block.

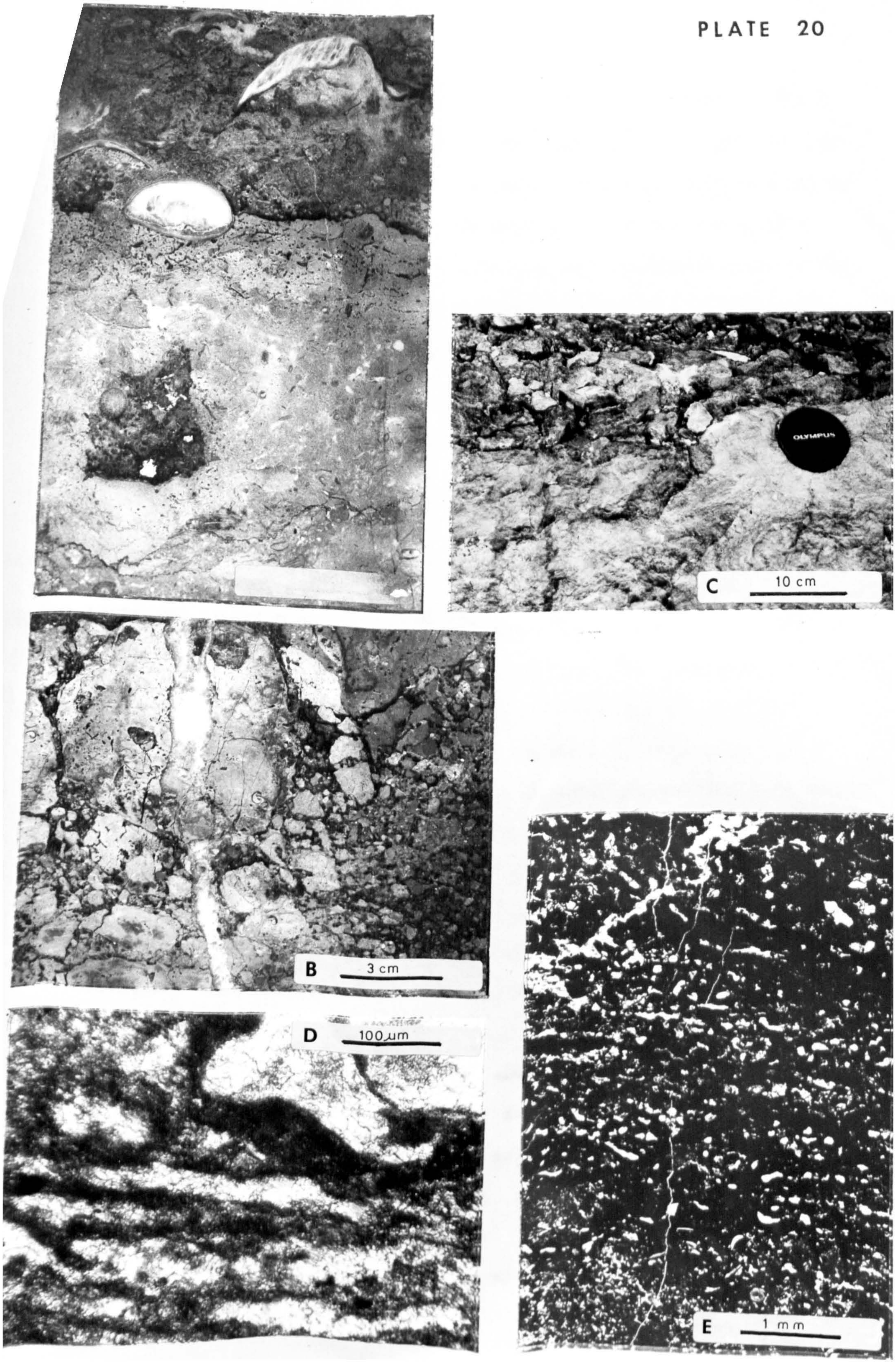
B Autobreccia with clasts coated by dense laminated calcrete, mature profile, Brigantian, Graig Quarry (Chart D)

C Neptunian "fissure" of calcrete underlying a laterally persistent laminated calcrete crust, loose block Bron Heulog Quarry ?from Eg. 8., Eglwyseg Formation.

D Fine micrite laminae within the millimetre laminae of a porous laminated crust, s/m 1313 G. R. SJ 221445

E Porous laminated crust with tubular structures, infilled by alveolar structures, s/m 1313 G. R. SJ221445.

PLATE 20



impregnation of calcium carbonate within host 'soils' s.l. First used by Lamplugh (1902), and subsequently interchanged with 'caliche' the term 'calcrete' has now become readily acceptable for many $CaCO_3$ precipitation fabrics and textures associated with soil formation reflecting variation in climate, drainage, host rocks, soil chemistry and maturity of processes. (Reviewed in Goudie, 1973)

Within the Asbian and Brigantian succession of the Llangollen and Oswestry 'embayments' calcrete profiles are of two related forms: rarely as mature (deep) brecciated and coated profiles and more commonly as superficial profiles with laminated calcrete crusts and/or tubular calcrete structures.

11.3.1. Petrographic Description.

11.3.1.1. LAMINATED CALCRETE CRUSTS. Laminated calcrete crusts† occur as thin light ochre to deep brown finely laminated horizons (between 1mm and 30cm thick) located towards the top of minor cycles, commonly immediately underly-ing palaeokarstic surfaces, or rarely within monotonous sequences of bioclastic and peloidal grainstones (see chart E). The crust laminae follow underlying natural surfaces within or upon the sediment. They are wavy 1 to 3mm thick, dark/light couplets, consisting of even finer micrite laminae. These finer laminae, also concentrically coat interstitial peloids and clear calcite spar-filled 'fenestral' cavities, that locally account for up to 50% of more porous crusts. Visible in very thin section, the finer micritic laminae are about 20μm, at times interlayered with thin lensoid and pervading calcite spar 'voids' of laminoid fenestral form (Plate. 20 fig. D). 'Packing' fenestrae pervade the larger laminae, defining peloids and ?peloid aggregations. These peloids are homogeneous micrite, rarely greater than 250μm diameter, and have rounded margins.

Deeper rust coloured laminae within the crusts are rich in ?iron

† Plate 20

oxide (?after pyrite framboids), and fenestrae have ?early isopachous brown ferroan calcite fringes up to 50μm thick, with ?iron carbonate inclusions, abruptly terminating against clear void filling druse.

The fenestral structure of crusts is dominated by a tubular element†, with both vertical and more common horizontal components. These larger fenestral features are pervaded by 'alveolar' micrite partitions (see on). Some thin crusts (e.g. s/m 621, Trefor Formation, Trefor Rocks, G. R. SJ.23024360) have distinct zonation from "non-porous" (unfenestrated) to "porous" with a high degree of fenestration.

The surrounding sediment mostly grades into the crust via intensely micritised 'peloidal' grainstones with alveolar textures and thin micrite laminae enveloping,pervading (?incipient pisolite structures) and adjoining adjacent clasts. Clasts are micritised inward from the periphery, and this process is most complete with clasts incorporated within crust laminae. Thin discontinuous (1 - 10cm thick) sediment layers rarely pervade crust profiles, giving a'multiple' crust form, e.g. top of the second Brigantian cycle, Craig Quarry, Chart D.

Insoluble residues of crusts are dark grey to black (formic acid extraction). During dissolution an 'oily' scum of amorphous kerogen (J.E.A. Marshall, verbal communication) forms on the solvent surface. X.R.D. analysis of the ≤ 2μm fraction of the residues show that, although mixed layer illite/smectite and smectite peaks are comparatively weak. Kaolinite 001 reflections are relatively strong to moderate, similar to suprapalaeokarst palaeosols (Appendix III)

2.3.1.2. <u>ALVEOLAR TEXTURES</u>. Within primary porosity associated with calcrete crusts, calcretised tubules and associated sediment, alveolar structures are commonly present, distinguished as thin (10 to 50μm) <u>partitions forming</u> a vesicular and reticulate pattern in cross-section

† Plate 20, fig. E

(Plate 22 fig. A). Partitions are generally curved, and are composed of micrite laminae of similar grain size (1 - 3µm) and character to the calcrete micrite of crusts and tubule walls. These partitions define chambers rarely in excess of 500µm diameter, and more commonly between 200 and 300µm. The chambers are at times perfectly spherical, comprising touching spheres, with diamond shaped (in cross section) voids between, but more commonly highly irregular in shape. Clear spar druse infills the partitioned voids (Plate 22, fig. B).

Large horizontally elongated 'clots' of alveolar texture occasionally occur within sub-palaeokarstic sediments (e.g. s/m 1531) up to 10 by 4mm.† In s/m 2004 (World's End G. R. SJ. 23654767) clasts of syringoporoids with well developed micrite rims have been partly dissolved, leaving the micrite rim as a template (Bathurst, 1966), into which alveolar partitions have penetrated . The alveolar networks do not, however, penetrate adjacent mollusc clasts that have also undergone ?void dissolution (Bathurst 1966). In MA.3 sediment adjacent to the alveolar void systems, alteration may be negligible or thin micrite veneers may coat clasts and fill intergranular voids.

11.3.1.3. <u>TUBULAR CALCRETE STRUCTURES.</u> Superficially similar to tubular fenestrae (section 6. 4.2.3) these penecontemporary voids have dark brown calcrete micrite walls‡, varying up to 400µm in thickness with a fine concentric lamellar structure (Plate 21 fig. C, D). They incorporate and coat adjacent sediment clasts within the walls which diffuse into adjacent sediment (Plate 21 fig. F). These tubules rarely cross cut alveolar 'clots', and are dominant components of laminated calcrete crusts. They themselves are often partitioned by an alveolar fabric. Often present as the only calcrete structure beneath palaeokarsts, they may extend as horizons up to 1m thick into the underlying sediment (e.g. fig. 25.4).

† Plate 22 fig. A . ‡ Plate 21, fig. B

PLATE 21 Tubular calcrete structures

A Tubular calcrete structures, with much admixing of clays (black). S/m 1310, Eglwyseg Formation,'Eg. 4',Dinbren Uchaf G. R. SJ 22004459.

B Calcrete micrite wall of tubular calcrete structure, s/m 1310, Eglwyseg Formation.

C & D S.E.M. photomicrographs. Laminar nature of calcrete micrite tubule walls. S/m 1310 Eglwyseg Formation. 10% HCl etch for 60 sec.

E Large tubular structure associated with calcrete profile

F Polished slab: Tubular calcrete structures. Note diffuse outer margins to dark micrite walls, and geopetal silt infills (pale). S/m 1310 Eglwyseg Formation.

PLATE 21

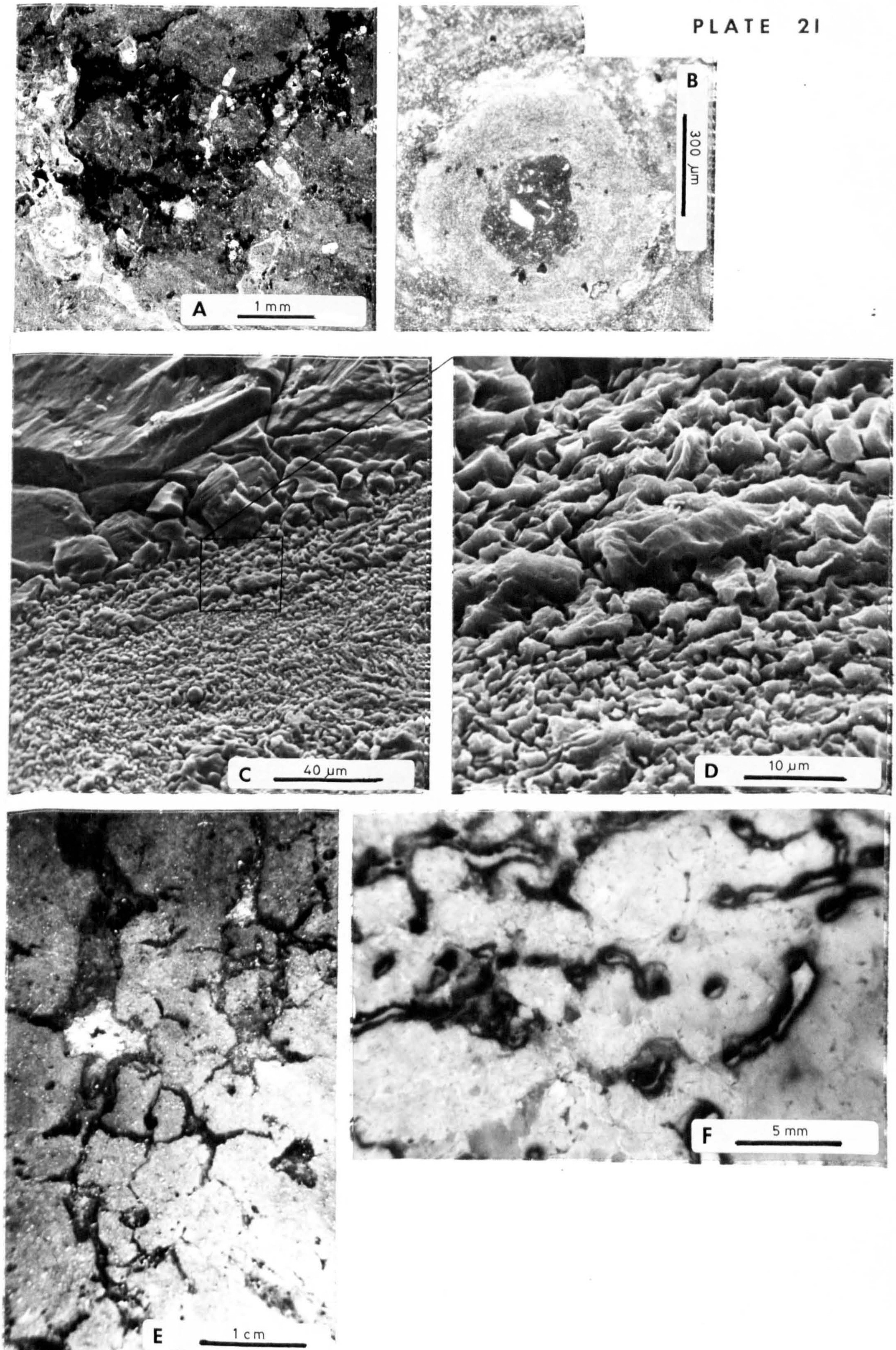


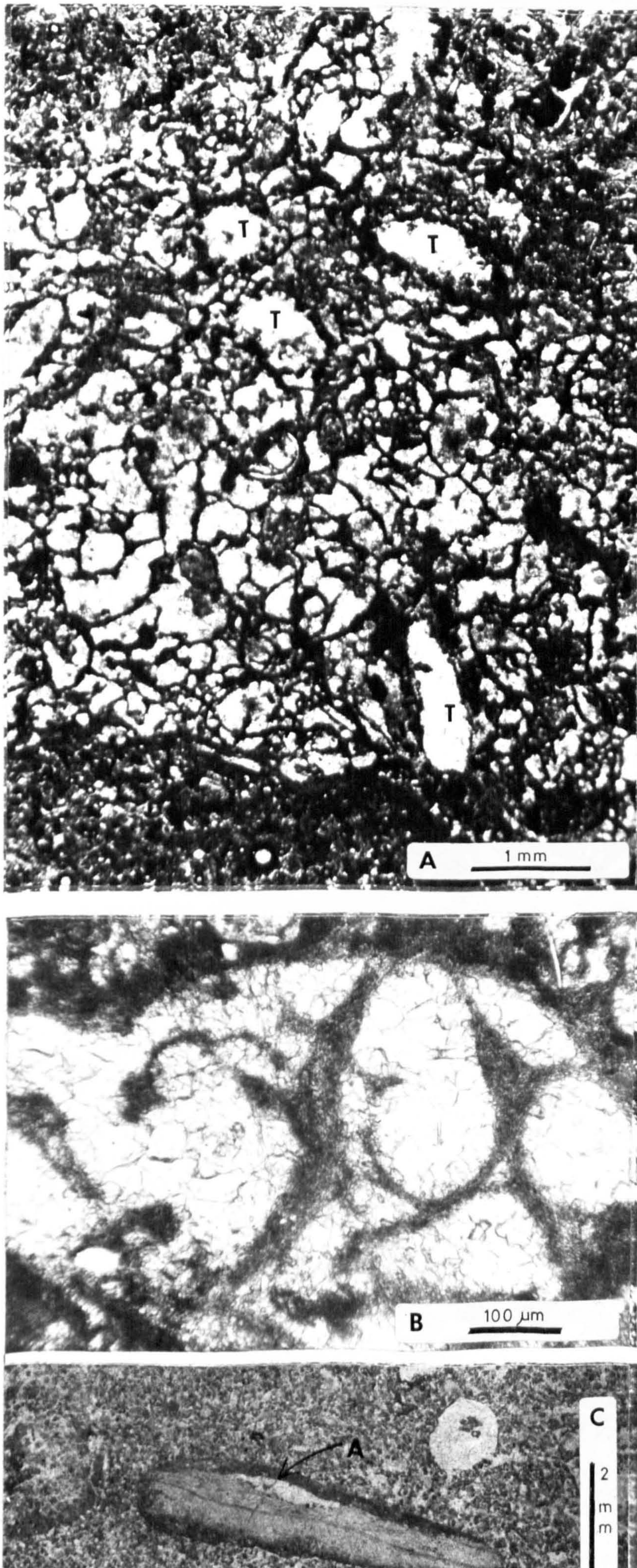
PLATE 22

A Large 'clot' of alveolar texture, transected by tubular structures. Note apparent thickening of tubule wall of. alveolar 'wall probably apparent refl- ecting push-apart formation of the tubules. S/m 1531 Mesothem D5b, Loggerheads Formation, Erryrys (Chart B

B Alveolar texture. Note roundness of chambers, vague laminar microstructure to calcrete micrite walls, and small packing triangles etc. between the subspher- ical chambers. S/m 2004, Eglwyseg Formation, World's End, G. R. SJ 23654767.

C Alveolar texture (A) penetrating dissolution feature within a micrite envelope of ?syringoporoid clast. S/m 2004 Eglwyseg Formation, World's End.

PLATE 22



Sediment grains appear more tightly packed around the tubule structures in grain supported sediments. Although this effect may, in part, be due to the presence of calcrete micrite around the tubule, it indicates that the tubules formed by push-apart rather than active dissolution processes (no grains have been observed truncated), and that the sediment was, therefore, at most, only partly lithified.

The size variation of tubules is not great. Rare examples exceed 5mm diameter(including the wall)† but most lie within the range 800 to 400 μm. Internal diameters are even less variable, mostly about 300 - 100μm.

The interior of the tubules may be infilled by spar druse, alveolar texture, crystal silts or, in more lime-mud lithologies, the surrounding sediment where they have typical fenestral infill patterns (section 6. 4. 6).

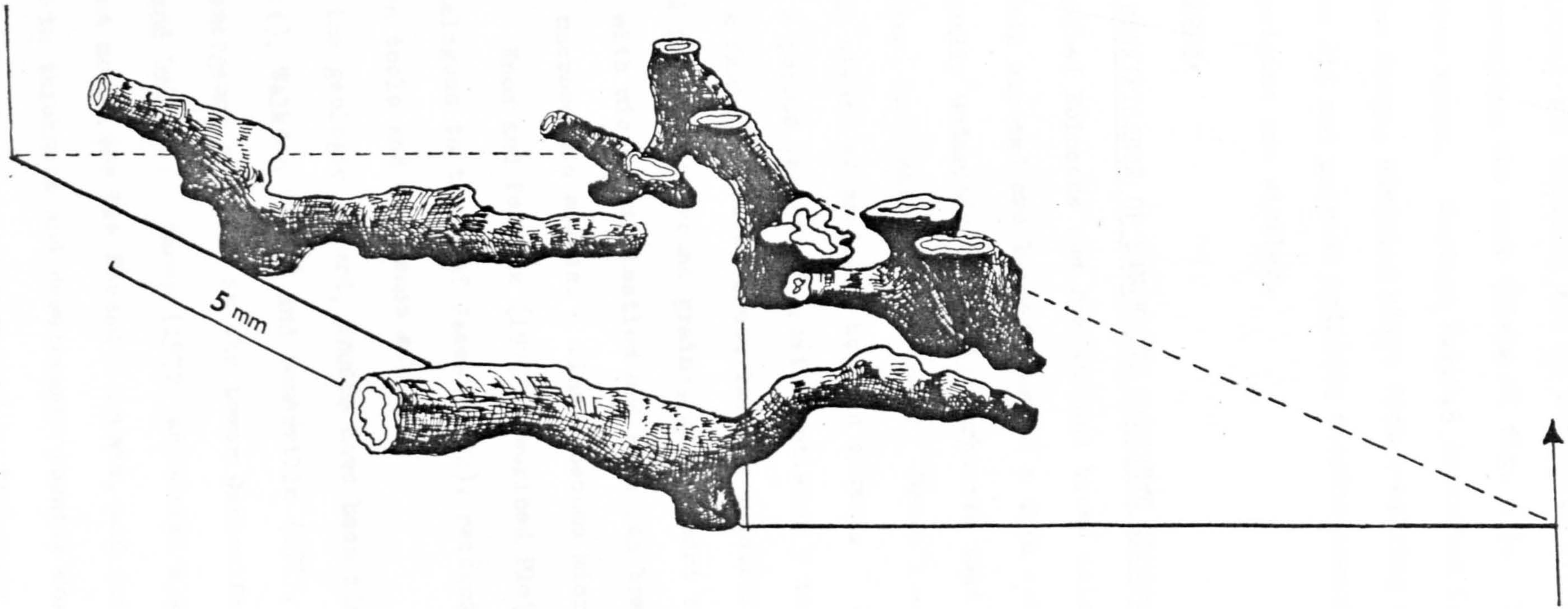
Serial acetate peels of an orientated specimen indicate that the tubules branch irregularly, (fig.73), but maintain constant diameters during branching.

The tubule systems are dominated by horizontal components (cf..more vertical tubular fenestrae of section 6.4.2.3). Vertical links are present, but less common, and some individual tubules can be traced for greater than 5cm laterally. Significantly, these are the only calcrete structures preserved along many ?cycle boundaries or palaeokarstic surfaces, apparently in part due to palaeokarstic removal of overlying crusts. In well developed crust profiles (e.g. boundary Eg.2a, Trefor Rocks, Eglwyseg Formation) a zone of unaltered sediment underlies the laminated crust, ramified by these tubular calcrete structures. (Plate 20 fig. A is an exception)

In acetate peels of more micritic sediments containing tubular calcrete structures (e.g. s/m 1310), clays are visible, both concentrated

† Plate 21, fig. E

TUBULAR CALCRETE STRUCTURES



RECONSTRUCTION OF TUBULAR BRANCHING PATTERN FROM SERIAL ACETATE PEELS within and around the tubules, and as irregular stringers up to 5mm diameter, permeating the rock (Plate 21 fig. A). Stylolites tend to follow these zones. Section 6.4.1.3 described irregular tubules of a similar size range, admixing clays from overlying sediments. Although these tubules did not possess calcrete micrite sheaths, it is probable that their origins are similar.

11.3.2. Interpretations.

11.3.2.1. ORIGIN AND SIGNIFICANCE OF LAMINATED CALCRETE CRUSTS. James (1972) described Holocene and Pleistocene brown calcareous crusts from Barbados, both exposed and buried beneath a thin soil cover. They are individually underlain by chalky carbonate that grades down, beneath the crust zone, into marine limestones. Brown 1 - 3μm micrite micro-laminae coat clasts of sediment between crusts. James (op.cit.) recognised a porous laminated crust comprising 5 to 20% horizontal root tubes, and a dense laminated crust lacking quantities of root tubes. James noted fracturing around grains, reminiscent of packing fenestrae, associated with microsparitisation and solution brecciation on both micro- and macroscopic scales. Clasts become micritised during crust formation. Enos and Perkins (1977) described Pleistocene crusts from Florida analogous to those of James (1972), recognising them as useful correlation tools and emergence events.

In the geologic record, crusts have been identified by Walls et.al.(1974), Walkden (1974) and Somerville (1977, 1979a, c) on cycle boundary emergence surfaces in the Lower Carboniferous, both in the U. S. A. and Britain. James (1972) advocated a grain growth and replacement model for his Barbados crusts, and noted that the custs increased in thickness and development towards coastal spray zones, that add $CaCO_3$ to nearsurface waters. Carbonate taken into solution, in James'(1972) model, is reprecipitated during periods of intense evaporation i.e. formation of calcrete crusts require neither highly pluvial nor arid environments. Recent cryptocrystalline brown calcrete is low-magnesian calcite. Robbin and Stipp (1979) estimated growth rates of Holocene crusts from 1cm/2 000 to 4 000 years, showing that they are accretionary rather than due to a descending zone of recrystallisation.

The biochemical influences in development of laminated calcrete crusts are prominent. Walkden (1974) noted that heating blackened the crusts,indicating high carbon content and Klappa (1979) recognised indurated lichen stromatolites as incipient laminated crusts on exposed calcrete hardpans from the Mediterranian. Braithwaite (1975, p. 21) noted fungae and filamentous algae may have influenced deposition, but recognised their accretionary formation rather than in situ modification of existing sediment.

Whether or not Lower Carboniferous laminated calcrete crusts formed with the aid of biochemical carbonate precipitations, the effect of evaporation-precipitation in forming the calcrete is apparently most significant, by comparisons with Recent and Pleistocene crusts, with fine micritic laminae reflecting a cyclicity in this process. Crusts 15cm thick (e.g. boundary Eg.2a, Trefor Rocks), using Robbin and Stipp's (1979 formation rates required about 30 000 - 60 000 years to form. This is a length of time comparable in duration to that estimated for formation of palaeokarstic surfaces of low to moderate relief (Form A).

11.3.2.2. ORIGIN AND SIGNIFICANCE OF TUBULAR CALCRETE STRUCTURES. Enos and Perkins (1977, p.143) described calcified roots and root tubules in Floridan Pleistocene laminated calcrete crust profiles. Such root tubules are commonly thin (1 - 3mm diameter (James 1972, p.825)). They occur both within Holocene and Pleistocene crusts and underlying porous zones, "penetrating soft chalky carbonate in profusion". Enos and Perkins (1977) reported larger calcified 'tap-root systems', orientated vertically. Significantly they noted that locally the sediment was composed almost exclusively of these tubules and their calcrete micrite walls, as appears common in some of the Asbian crusts encountered. Roots have been observed intact in some of these late Holocene crusts, although Pleistocene examples retain root pore space only, after decomposition. Enos and Perkins (1977) could not recognise the plant(s) responsible, but Hoffmeister and Multer (1965) attributed some Holocene calcified root structures to the black mangrove *Avicennia nitida*. Walis *et al.* (1974, p.429) compared similar Carboniferous structures with worm tubes, but supported a root origin. Read (1974b) also described calcrete zone root casts from Quaternary sediments of Shark Bay. Braithwaite (1975, p.9) included calcretised root tubules from Aldabran sediments as 'pedotubules', however according to Brewer (1964) "pedotubules" are filled tubular structures having a different plasmic structure to the surrounding sediments. The tubular structures described here are "channels" according to the Brewer (1964) scheme (except where rarely infilled by sediment).

The lack of tap root systems, and unobvious 'root hairs' imply that roots may not be responsible for their formation, but neither interconnecting tubules (in a reticulate manner) nor faecal pellets have been observed (cf. V.P. Wright, abstract of Palaeontological Association Annual Meeting, 1980) which may suggest an infaunal origin, as envisaged by Brewer(1964) for many 'channels' in soil profiles.

Braithwaite (1975) considered the dense micrite wall of his tubules to be of "calcitan"(Brewer 1964) cutanic origin. As cutans are linings on natural surfaces within soils, then many laminar calcrete phenomena may be considered 'cutanic', and replacive calcrete fabrics

The morphology of the vesicles within these Asbian alveolar masses is consistent with entrapped air bubbles, locally well formed and not interpenetrating, but mostly highly packed, giving polygonal sections. 'Space' between the air bubbles (small 'diamond' shapes†), and the bubble interfaces themselves, filled or veneered with carbonate saturated solutions would precipitate calcrete micrite in thin laminar coatings, on the bubble surface during desiccation as a 'meniscus' structure (Dunham, 1971).

Once initiated, the calcareous envelope may have acted as a template for subsequent bubble formation. Presence of air pockets within the sediment is consistent with a vadose environment (as predicted for calcrete formation). The bubbles may have formed during periods of (solar) heating following intermittent pluvial events, with increase in temperature inducing $CaCO_3$ precipitation.

11.4. PALAEOSOLS.

Overlying some palaeokarstic surfaces (and more rarely calcrete profiles), filling depressions within the irregular surfaces, are unfossiliferous coloured, and rarely laminated claystones. They vary from greys to maroons, greens and ochres, and occasionally stain the underlying carbonate. They may grade upwards into fossiliferous grey shales (e.g. on boundary Eg. 3, Trefor Rocks) or intraformational and extraformational conglomerates.

Walkden (1972) and Somerville (1977) analysed the clay mineralogy and geochemistry of these claystones. They have a high potassium content (4 – 8% K_2O), high Sr and Rb and low free silica with a zircon apatite and anatase heavy mineral suite (Somerville, 1979a, p.323). Their clay mineralogy is variable but includes mixed layer illite/smectites, smectites and kaolinite. Ap'ix III shows their clay content variation. In comparison clays from interbeds within minor cycle

† Plate 22 fig. B .

basal phases and insoluble residues of various lithologies are illite dominated. Comparisons of clay mineral proportions were estimated by measuring 001 reflection peak areas from diffractograms, and expressing their areas as relative ratios. This comparison of clay component ratios eliminates, in part, effects of sample volume, grain size and orientation variations. Whilst mixed layer illite/smectites + kaolinite + smectite dominate clay assemblages in coloured claystones overlying palaeokarsts, their sum (of 001 peak areas) is less than the illite component in inter-bedded (non-palaeosol) shales (Ap'ix III).

Both Walkden (1972) and Somerville (1979a, c) interpreted palaeosols as having a wind-blown volcanic origin, although Walkden (1970) recognised that insoluble residues from the underlying karstified limestones also contributed to the clay assemblage.

11.5. PROFILE DEVELOPMENT

Only one of 35 sites of well exposed subaerial emergence surfaces in the Eglwyseg Formation (table 9) shows the development of palaeokarst, calcrete and palaeosol (boundary Eg.2, Tynant). The deeper palaeokarstic surface reliefs (Forms B and C) do not possess associated calcrete structures, and most Form A surfaces possess calcrete structures only intermittently. Crust laminae are commonly truncated by the palaeokarstic surfaces, and only one has been observed where the crust profile follows an undulose ?palaeokarstic surface (Eg. 2a, Trefor Rocks) but even in this case up to 50cm overlying the calcrete a non calcretised form A palaeokarst occurs. Calcretisation in most† cases, therefore predates karstification (see section 11.6.)

Calcrete structures are very rare in the Whitehaven Formation, where Form A and D palaeokarsts are common with overlying palaeosols.

† None have been observed otherwise.

TABLE 9

FREQUENCY OF EXPOSURE OF ASBIAN EMERGENCE PROFILES.

Formation	Palaeokarst only	Palaeokarst + tubular calcrete structures	Pkarst + tub. calc. struct. + laminated calcrete crust.	Pkarst + palaeosol	Pkarst + calcrete + p'sol.	Calcrete only	Total
Eglwyseg Formation	10	2	6	8	1	8	35
Sychtyn Member + Llynclys Form^{n}.	12	0	0	8	0	0	20
Llanymynech Member	7	0	1	3	0	0	11
Kingsdale Limestone (Ingleton)	0	2	0	4	1	0	7

In the Tynant Formation, palaeokarsts are difficult to identify (section 6. 3), and most cycle boundaries are planar ?erosion surfaces or S. D. surfaces.

At Graig Quarry (G. R. SJ 204567) a very mature calcrete profile has developed on the top of Somerville's (1979c) Cefn Mawr Limestone Formation cycle'l'(Brigantian), chart D. At this locality, a double laminated crust (separated by 8cm of bioclastic peloidal grainstone), up to 40cm in total, is developed, but the lower dense crust has incorporated and coated on to brecciated fragments of underlying calcrete micritised carbonates, from coarse sand to 10cm(+) in diameter, (Plate 20 fig. B). Small neptunian vertical 'fissures' are infilled with dense laminated crust, laminae paralleling the fissure walls and extending up to 20cm into the underlying rock. Above this profile, only a low 5 - 20cm amplitude (Form A) palaeokarst is developed.

A mature calcrete profile is also developed on 'boundary Eg.8' Eglwyseg Formation, at Bron Heulog Quarry but here a 20+cm laminated crust , with a neptunian and irregular lower surface (Plate 20, fig. C) is underlain by a zone, 1m thick, pervaded by tubular calcrete structures.

Insoluble residues from calcrete profiles, both laminated crusts and tubular calcrete structures, contain a clay mineral assemblage more similar to palaeosols than non-calcretised sediments, with kaolinite, mixed layer illite/smectite and smectite. Appendix III illustrates X-ray diffractograms of the palaeosols and calcrete insoluble residues. The formation of tubular calcrete structures, and disruption of surrounding sediment by biomechanical processes may have caused a mixing of soil clays with underlying carbonate, especially if the carbonate was only partly lithified, as indicated by the tubular calcrete structures themselves.

If laminated calcrete crusts are in part biochemically formed, and analogous to Klappa's (1979) lichen stromatolites, they grew on exposed rock platforms, and during growth were at most only temporarily covered by palaeosol. The presence of sediment layers within dense and porous crusts implies that short lived sedimentation events occurred upon calcrete surfaces, either with no overlying soil or after deflation of a soil cover. Neither soil clay whisps nor laminae are interlaminated with the crusts, nor is there any indication that the crusts formed within a clay soil. Their insoluble residues are less than or equal to those (in quantity) of underlying pure carbonates, and possibly represent a very small supply of wind-blown soil clays trapped by surficial vegetation. Fig. 74 illustrates the most likely model for profile formation with the eventual development of a palaeosol, both retarding calcrete formation, and enhancing palaeokarstic dissolution, that in its extreme, will remove previously developed calcrete features. This may be palaeoclimatically controlled (section 11.6.)

Palaeokarstic dissolution may, at times, have commenced by modification of an S. D. surface. Irregular, but palaeokarstic relief, illustrated in fig. 69 on an emergence surface in the Sychtyn Member may have evolved from an initial S. D. surface. Reasons for scarcity of such features are discussed in section 11.1.3.

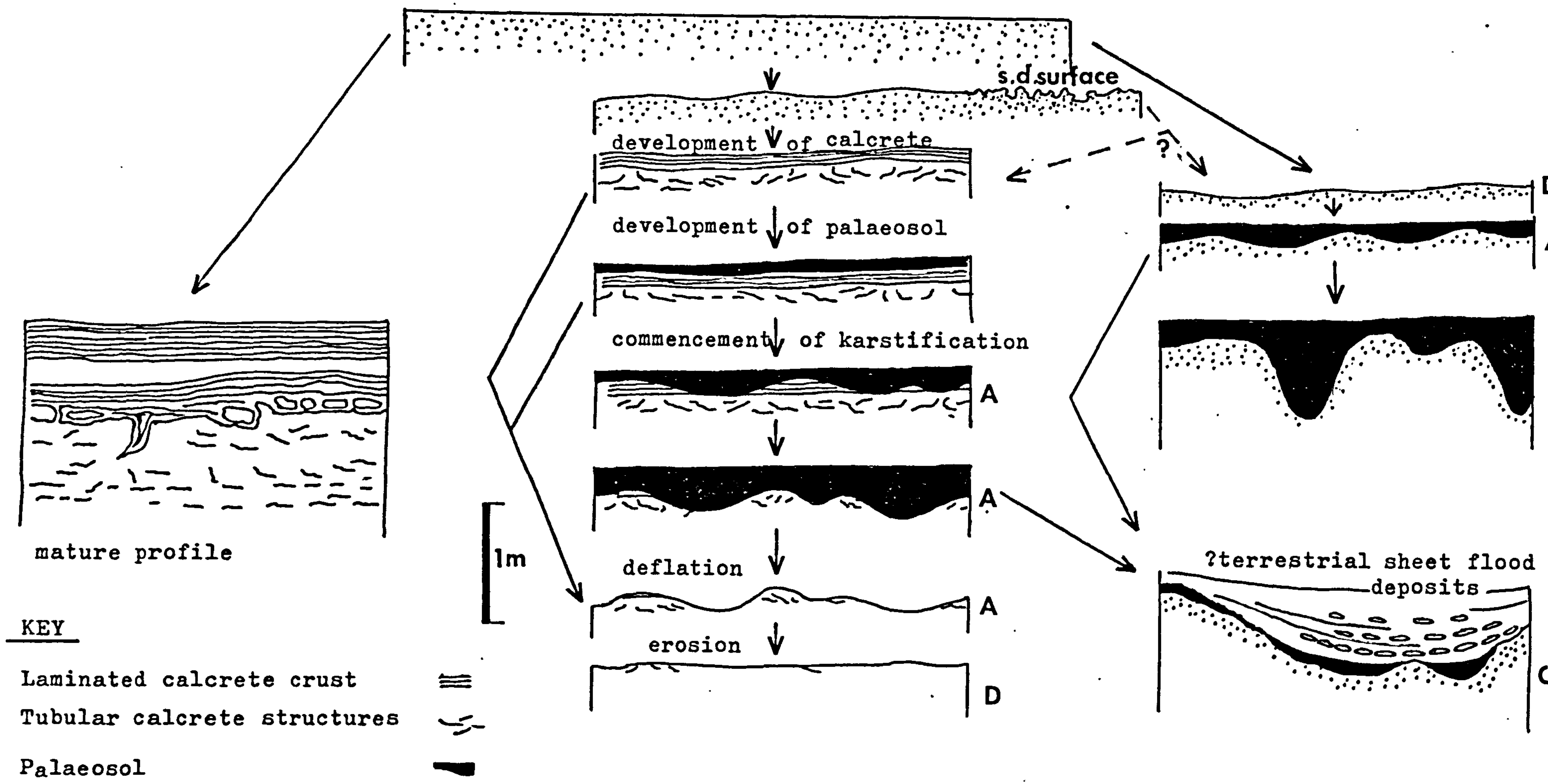
Once intergranular cementation has started, enhanced through karstic dissolution by meteoric waters, capillary transport and calcrete micritisation will become stifled.

Floridan (Pleistocene) crusts (Enos and Perkins, 1977) line solution 'pipes' indicating growth after or during dissolution.

Deflation and erosion associated with succeeding transgressions modified some surfaces producing near planar topographies, and removing palaeosols.

In the genetic sequence of profile development (fig 74) it is noteworthy that the observed number of emergence surfaces at each stage

SUBAERIAL EMERGENCE PROFILE GENESIS



Arrows indicate probable sequential relationships of emergence profiles

Fig 74

Letters to right of diagram indicate palaeokarst form (see text).

of development (in the Eglwyseg Formation) reduce with maturity. Shorter emergence events were apparently most common, although dependent upon the rate of regression a single emergence surface may have a more mature profile towards more proximal shelf areas. Fig.75 shows the lateral variations along Eglwyseg Formation palaeokarsts (the better exposed ones). Although the data is limited, boundaries Eg.2, Eg.4, Eg.5, Eg.8, and Eg.9 show some degree of increasing maturity (according to the scheme outlined in fig.74) southwards between Llwyn Hen Parc and Bron Heulog.

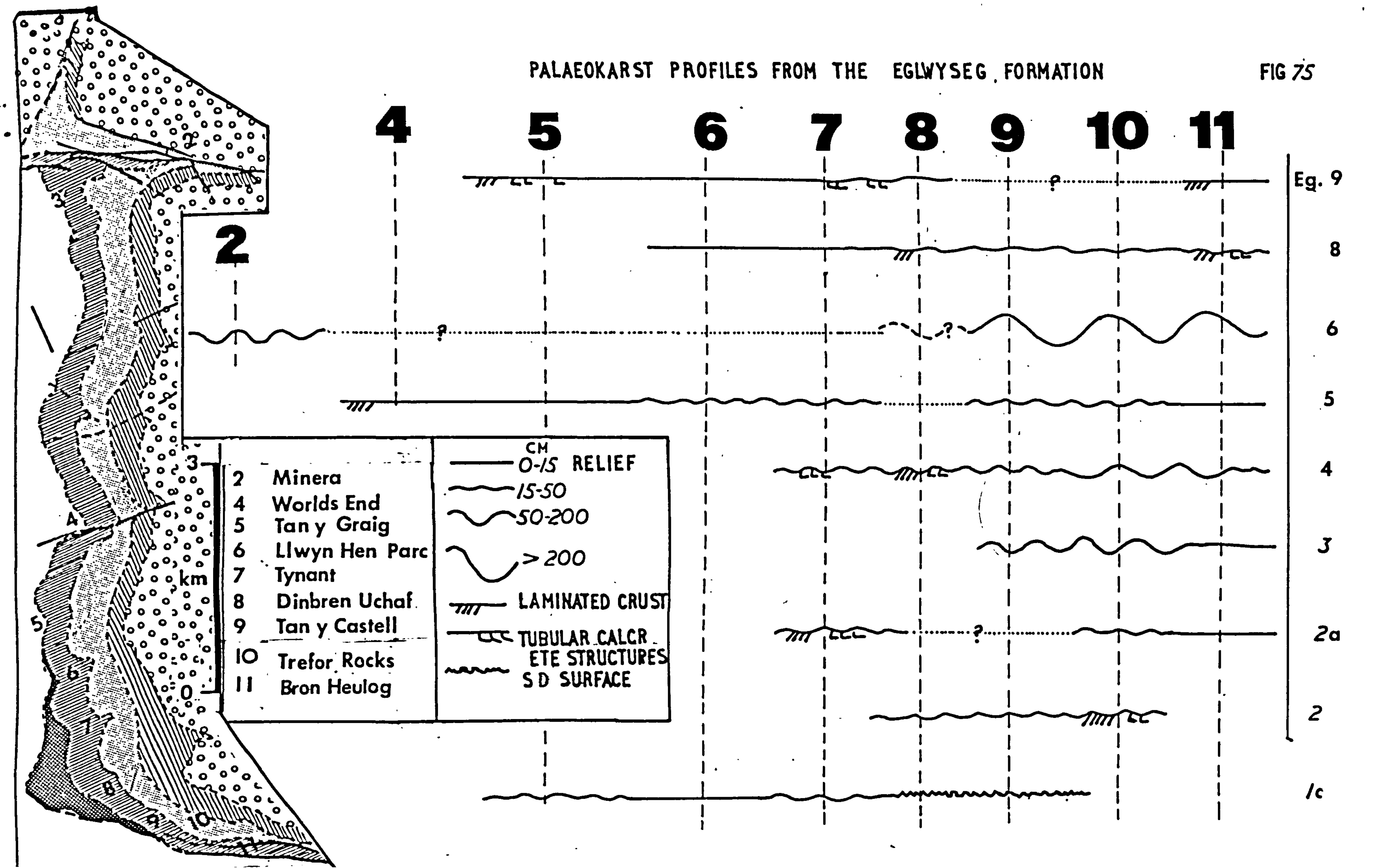
Intra- and extraformational conglomerates infill palaeokarstic surfaces with deep reliefs (2-5m; Form C) along Trefor Rocks, between Tan y Castell and Bron Heulog (Eglwyseg Formation). These conglomerates lack any internal structure except crude lamination, and reflect periods of intense erosion on the emergent shelves and Lower Palaeozoic hinterland during periods of storm and sheet floods.

Within 2km of these Form C palaeokarsts is Dinas Bran, a prominent conical hill of Ludlovian litharenites. This may be a relic of a local positive influence to Carboniferous emergence of the Llangollen Embayment.

11.6. PALAEOCLIMATIC CYCLICITY

Climatic variation during the Pleistocene has been well documented. Holmes (1965, p.702-703) reproduced a palaeoclimate fluctuation curve showing ?composite cyclicity with larger cycles of 100 000 to 300 000 years duration corresponding to the major glacials and interglacials. Whether similar (more subdued) fluctuations occurred in the Lower Carboniferous (for which there is no positive evidence of an ice cap) is unclear. (Chapter 12).

Palaeokarstification of emergence surfaces tends to occur after calcretisation, and overlying palaeosols lack calcrete nodules and


PALAEOKARST PROFILES FROM THE EGLWYSEG FORMATION
FIG 75
2
4
5
6
7
8
9
10
11
Eg. 9
8
6
5
4
3
2a
2
1c
2 Minera
4 Worlds End
5 Tan y Graig
6 Llwyn Hen Parc
7 Tynant
8 Dinbren Uchaf
9 Tan y Castell
10 Trefor Rocks
11 Bron Heulog
CM
0-15 RELIEF
15-50
50-200
> 200
LAMINATED CRUST
TUBULAR CALCRETE STRUCTURES
S D SURFACE
km
3
0
A

glaebules etc.. Whilst karstification reflects relatively pluvial palaeoclimatic conditions, the development of calcretes requires a more arid and evaporative palaeoclimate (a few calcrete profiles extend down metres within the underlying sediment, an indication of the depth of the water table during their formation.). Unfortunately, little evidence is available for distinguishing palaeoclimatic fluctuation during the deposition of the carbonates themselves.

This implies that a change to a more pluvial palaeoclimate may have occurred during emergence phases. This calcrete-karst transformation may otherwise reflect eventual development of a soil profile.

Gradual deepening of a water table, associated with a continued regression (and relative uplift of the shelf) would not cause karst initiation.

Rognon (1980) described pluvial and interpluvial periods during the Pleistocene of equatorial Africa. Similar events have been recorded in more temporate latitudes (Holmes, 1965). Pluvial periods, characterised relative to Recent climates, were believed by earlier workers (in Bowen, 1978) to correspond to glacials (with world-wide sea level lows) and interpluvial periods with interglacials (with world-wide sea-level highs). Bowen (1978) however, referring to the African Pleistocene situation, indicated that the classification of pluvial/interpluvial phases was a constraint to world-wide and local correlations, and that their local recognition was a consequence of both local (principally) and world-wide climatic factors.

Rognon and Williams (1977), however, described coexisting pluvial and interpluvial episodes in the late Pleistocene from Africa and Australia. They related interpluvial episodes to periods of persistent anticyclone development, that reflected positioning of subtropical jet-streams. Rognon (1980) demonstrated (for North Africa) the effects of local geography on development of apparent pluvial episodes, readily seen as lake-levels fluctuations during the Pleistocene.

Stanley and Maldonado (1979) described palaeoclimate-eustatic sedimentary cycles from the Nile cone, but correlated these with major changes in palaeoclimate and eustacy, rather than complete 'minima to minima' cycles.

It is conceivable that the Lower Carboniferous climates reflected a global climatic cyclicity, also inducing minor glacial events (or were both controlled by another mechanism) in Gondwanaland, which produced glacio-eustatic fluctuations (section 1.2.1.1), with 'pluvial' events corresponding to the development of palaeokarst and sea level lowering. Conversely, the formation of calcretes and development of sea level maxima (and deposition of the carbonate sediments) may reflect periods of comparative aridity.

12. PALAEOGEOGRAPHIC CONTROLS, CYCLE MECHANISMS AND INTER-RELATIONSHIPS

Sedimentary cyclicity is the product of rhythmically migrating facies belts. The principle factors inducing this sedimentary cyclicity within the Lower Carboniferous are eustacy, tectonism and progradation. These three mechanisms are, however, intimately related as global sea level changes (eustacy) allow sediment progradation following transgression, both of which may at least in part influence local and or regional tectonism. Tectonism is a major influence upon palaeogeography which may be readily related to cycle patterns and variation.

12.1. MAJOR MECHANISMS CONTRIBUTING TO SEDIMENTARY CYCLICITY IN THE LOWER CARBONIFEROUS

12.1.1. Eustacy.

Eustacy is probably the most important single cyclic process acting on sedimentation during the Lower Carboniferous in North Wales. Eustatic sea level changes can be induced primarily by either ice-cap fluctuations (fluctuations in the amount of the Earth's surface waters constrained as continental ice) or by changes in the cubic capacity of ocean basins as a result of widespread ocean floor tectonics (Duff et al 1967, p.245).

Donovan and Jones (1979) reviewed the potential of eustatic processes to provide geologically significant sea-level changes. They calculated that a total melting of existing continental ice-caps would account for a 40-50m rise in global sea-level. They also concluded that, during the Pleistocene, the maximum sea-level fall due to growth of continental ice-caps was about 100m. This correlates closely with observed geological data (e.g. James and Ginsburg, 1980, p.168). Clearly in the Asbian and Brigantian transgression or regression on this scale is neither acceptable nor required.

Donovan and Jones (1979) concluded that eustatic effects controlled by desiccation of ocean basins, changes in mean ocean temperature, and variations in atmospheric moisture content could only provide a sea-level change orders of magnitude less than by glaciations. They also noted that variations in sea-level due to changes in ocean-ridge volumes, although volumetrically significant could only produce a change rate of 1cm/1000 years, three orders of magnitude slower than glacioeustatic changes.

Ramsbottom (1973a, 1977, 1979) proposed an eustatic model for cycle formation in the Dinantian, and applied this model to NW Europe (Ramsbottom, 1979).

Boucot and Gray (1979, p.477) described the global Lower Carboniferous as a time of "lowered climatic gradient." According to Frakes (1979) the occurrence of ice caps is in part related to the positioning of poles over continental masses. He (op.cit. p.150-153) sited the early Carboniferous pole (from palaeomagnetic data) over southern Africa. There is no direct evidence of an ice-cap in the early Carboniferous (although both Devonian and Namurian glacial deposits have been described in South America (in Frakes, 1979). The stratigraphic positioning of early Carboniferous glacial sediments in the Rio Blanco Basin, Argentina, reported by Frakes and Crowell (1969) has now been refuted (Frakes, 1979, p.138). Upper Palaeozoic glaciations followed the 'polar fall due to growth of continental ice-caps was about 100m. This correlates closely with observed geological data (e.g. James and Ginsburg, 1980, p.168). Clearly in the Asbian and Brigantian transgression or regression on this scale is neither acceptable nor required.

Donovan and Jones (1979) concluded that eustatic effects controlled by desiccation of ocean basins, changes in mean ocean temperature, and variations in atmospheric moisture content could only provide a sea-level change orders of magnitude less than by glaciations. They also noted that variations in sea-level due to changes in ocean-ridge volumes, although volumetrically significant could only produce a change rate of 1cm/1000 years, three orders of magnitude slower than glacioeustatic changes.

Ramsbottom (1973a, 1977, 1979) proposed an eustatic model for cycle formation in the Dinantian, and applied this model to NW Europe (Ramsbottom, 1979).

Boucot and Gray (1979, p.477) described the global Lower Carboniferous as a time of "lowered climatic gradient." According to Frakes (1979) the occurrence of ice caps is in part related to the positioning of poles over continental masses. He (op.cit. p.150-153) sited the early Carboniferous pole (from palaeomagnetic data) over southern Africa. There is no direct evidence of an ice-cap in the early Carboniferous (although both Devonian and Namurian glacial deposits have been described in South America (in Frakes, 1979). The stratigraphic positioning of early Carboniferous glacial sediments in the Rio Blanco Basin, Argentina, reported by Frakes and Crowell (1969) has now been refuted (Frakes, 1979, p.138). Upper Palaeozoic glaciations followed the 'polar wandering' on its eastward path across Gondwanaland.

Whilst the Devonian was a period of global reduction in epi-continental seas, the Lower Carboniferous marked a period of widespread transgression, and probably reduced polar ice caps. The present lack of evidence of a Lower Carboniferous polar ice cap does not preclude its existence.

12.1.2. Local Tectonics.

George (1977) recognised the influence of local tectonics on Dinantian sedimentation. Bott and Johnson (1967) evoked isostatic equilibration (through mantle flow) for late Dinantian cyclic sedimentation, related to major tectonic hinges in Northern England.

The formation geometry within the Llangollen and Oswestry 'embayments' reflects the differential subsidence between the more positive Cyrn y Brain and Berwyn "axes", with more rapidly subsiding 'depocentres' near Llangollen and Oswestry.

As emergence horizons (and especially Sutured Discontinuity Surfaces) indicate the position of sea-level, relative subsidence of these axes may be calculated where whole sequences are correlated.

At Minera the Eglwyseg Formation (Egl to Tfl) is 56m thick.

At Tynant the Eglwyseg Formation (Egl to Tfl) is 148m thick. The upper and lower boundaries of the Eglwyseg Formation are major cycle (mesothem) boundaries marking regression/transgression events, and indicate similar base-levels (relative to sea level) for deposition of these boundaries. Therefore, the centre of the Llangollen 'embayment' subsided during deposition of the Eglwyseg Formation by 92m more than the Cyrn y Brain. Similarly, subsidence of the Llangollen embayment relative to the Berwyns is about 145m, and subsidence of the Oswestry embayment, with respect to the Berwyns is about 45m during the equivalent time period (corresponding to deposition of Mesothem D5b (Llynolys Formation).) According to radiometric data cited in George et al (1976, p.76/77) the Upper Asbian spans approximately 10m.y. (between 338± 4 and 328 ± 6). Ramsbottom (1977, p.284), however, estimated about 4m.y. for this mesothem.

George et al (1976, p.34) informally divided the Asbian at the equivalent position of the Tynant Formation/Eglwyseg Formation boundary. This boundary is readily correlated along the North Wales outcrop and marks a period of prolonged regressive (peritidal) sedimentation and minor cycle offlap. Ramsbottom (1977, 1979) defined the lower and upper Asbian individually as approximating to mesothems "D5a and D5b" respectively.

Assuming that the Eglwyseg Formation spans the 'Upper Asbian' of Mesothem D5b the relative subsidence of the Llangollen 'embayment' with respect to its southern flank may have approached about 40m/m.y. According to the correlation chart (fig. 22), within the Upper Asbian, subsidence was highest during deposition of the upper Eglwyseg Formation. There is a correlation between basal transgressive phase argillaceous and thin-bedded phases and areas of higher relative subsidence (section 3.1.3.1). Whether subsidence occurred as pulses in response to sediment loading (Bott and Johnson, 1967) or was a continued process, is unclear, but attenuation of these basal phases on to more positive areas may be a product of either faster relative subsidence rates (with respect to sea level) during early stages of a transgression, or as a consequence of the depositional gradient.

The Llynclys Formation is about 40m thick(maximum) and is the chronostratigraphic equivalent to the Eglwyseg Formation. It spans all of Ramsbottom's (1977) Mesothem D5b (approximating to the Upper Asbian). As the Oswestry 'embayment' was subject to the same eustatic influences as the Llangollen area, it is concluded that the embayment represents a more proximal shelf position than the Llangollen 'embayment'. The D5b Oswestry embayment was not, however, subject to the high relative subsidence of the Llangollen area, and offlap of minor cycles towards the end of the D5b mesothem affects this region more than distal shelf positions.

12.1.3. Regional Tectonics and Eustatic Trends.

Section 12.1.2 showed that relative tectonic subsidence can be defined over the deposition of the Eglwyseg Formation. This local subsidence may be complimented by regional subsidence and overall eustatic transgression (as evidenced from successive minor cycle and mesothem onlap).

The total effect of both tectonic influences and eustatic trends over the deposition of the Eglwyseg Formation is to raise sea-level relative to the base of the formation by the thickness of the formation. If the Cyrn y Brain was not undergoing uplift or subsidence relative to stable areas, regional subsidence and eustatic trends raised sea level relative to the base of the Eglwyseg Formation by 56m, and total subsidence + eustatic trend over the duration of a single cycle would raise sea level relative to the cycle base by the thickness of that cycle during its marine depositional phase.

12.1.4. Cumulative Sedimentation.

During deposition of a full cycle, sedimentation by progradation displaces its cumulative thickness of water depth on the shelf relative to the cycle base, and is a counter process to 12.1.3. Estimation of Carboniferous sedimentation rates is prone to much error.

The Eglwyseg Formation, spanning the Upper Asbian, was deposited in about 4 to 10m.y. (probably lower than higher within this range). At Llangollen about 14 minor cycles were deposited during this time, each therefore spanning about 300 000 to 700 000 years on average. Ramsbottom (1979, p.152) suggested 500 000 y, and Walkden (1974) about 200 000 y. Considering the average thickness of these cycles (9m ± 5m) at Tynant and the time span of most subaerial emergence horizons, a figure approaching 300 000 y is most probable for the cycles.

Taking the Eglwyseg Formation as a whole, at Tynant where the Formation is thickest, the average deposition rate over the Upper Asbian was < 4cm/1000 y. Similar overall deposition rates are given for other ancient shelf carbonate sequences in Wilson (1975, p.15).

These rates contrast with the rates of deposition of Recent carbonates that mostly fall between 0.7 and 1.2 m /1000 y (Wilson, 1975, p.15; Enos and Perkins, 1977; Bathurst, 1975). Neumann and Land (1975)estimated rates of Recent Bahamian lime mud and sand deposition of 12cm/1000 y, but this was in an area subjected to much off-shelf transport, and was 1.5 to 3 times the rate of actual sediment production.

Rates of deposition of Carboniferous carbonates must be compared to the Recent with caution. Carboniferous carbonate production may have been episodic. Seibold (1980) recognised that bottom currents can produce non-erosion and non-deposition in recent environments, and may transport 'manufactured' sediments off-shelf

(Neumann and Land 1975). The presence of many mixed generation clasts (in massive-bedded MA.3 sediments , section 9.3. 1) implies that considerable reworking of surficial sediments occurred.

Allowing for a degree of sediment compaction and pressure solution (e.g. to 2/3 original thickness) average sedimentation rates for individual Eglwyseg Formation cycles (with a 50 000y subaerial emergence surface developed thereon) is 10cm/1000y. Depositional breaks (e.g. bedding planes, and periods of argillite sedimentation on transgressive events may increase the purer carbonate sedimentation rates towards those observed in the Recent, especially for the upper massive-bedded MA.3 and peritidal phases.

Sedimentation, with facies progradation is an important factor influencing minor cycle development.

12.1.5. Sea level Fluctuations from Sedimentological Evidence (Eglwyseg Formation)

Basal phases of thin-bedded argillaceous alga packstones (MA.2) were apparently deposited below fair-weather wave base, but their algal assemblage suggests waters metres, rather than tens of metres deep (section 7. 5. 1), with beresellid carpets in part acting as sediment baffles, reducing any active erosion and fines removal. Succeeding bedded MA.3 (Lithofacies L.2) reflect a fluctuating carbonate-producing environment. With deposition below and above fair-weather wave base. Overlying massive bedded Lithofacies L.1 represents regressive phase sedimentation above fair-weather wave base, with sediment reworking, storm sedimentation, and rare development of bedforms towards the top of this phase, but lacking oolites. A widespread uniform environment is envisaged, in shallow waters 2 to 5m deep. The lack of peritidal sediments associated with these cycles suggests that regression was too rapid for their development. An estimated water depth fluctuation curve (relative to the plane of sedimentation) is drawn in fig. 76 .

Ramsbottom (1979, p.152), revising earlier opinions, believed that transgressive events of mesothems were slower than their regressions, but whether this follows for their contained 'cyclothems' is unclear. The regression associated with 'D5b' certainly appears to have been considerably slower than its transgression on to northern flanks of St. George's Land. If minor cycle transgression was slow the basal thin bedded clastic-rich sediments of the Eglwyseg Formation cycles may represent markedly slow deposition rates, in comparison to cycle upper phases. The lack of transgressive features suggests that transgression was too rapid for their development (cf. thin intraformational conglomerates on Tynant Formation cycle bases).

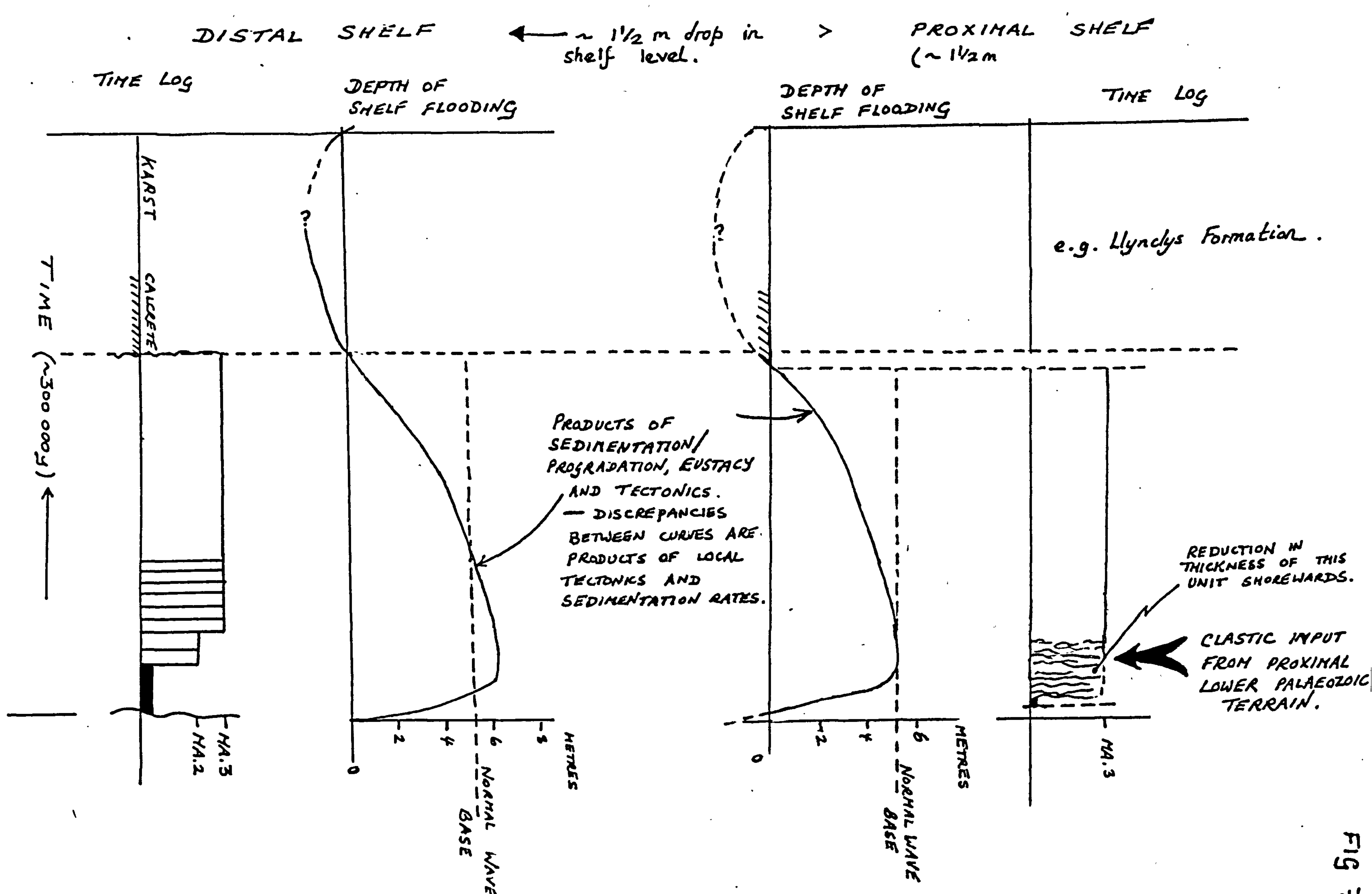
12.2. INTER-RELATIONSHIP OF CYCLE FORMS.

Significant observations on the Lower Carboniferous strata investigated in this study are the vertical and lateral relationship of the pervasive cyclicity.

The following observations are discussed:

1). Each chronostratigraphic unit comprises cyclicity of overall similar dimensions with characteristic trends.

2). Lateral variation in cycle form and stratigraphic packets reflects both local tectonic and palaeogeographic controls.



MODEL FOR LATERAL VARIATION IN MINOR CYCLICITY OF THE EGLWYSEG FORMATION & LLYNCLYS FORMATION

FIG 76

12.2.1. Chronostratigraphic continuity of Cycle Form

12.2.1.1. MESOTHEM D5a. The lower Asbian (Mesothem D5a) throughout the study area is characterised by a transgressive sequence of shoaling cycles that have marked peritidal suites as upper phases, with subtidal basal phases. They are, in general, thinner than overlying Upper Asbian cycles (e.g. as developed at Llangollen), but from evidence of their sedimentology they were deposited in a similar spectrum of environments. There is no evidence to suggest that relative sea-level fluctuations were more or less marked (in amplitude) for the minor cycles of Mesothem D5a than Mesothem D5b, but the frequency of each cycle type was apparently different.

In both the Llangollen and Oswestry depositional embayments transition from cycle type of Mesothem D5a to Mesothem D5b was a gradual process. The boundary in the Oswestry area is a prominent palaeokarst of about 50cm amplitude (form A). In the Llangollen embayment this same Mesothem D5a/D5b boundary varies from a sutured discontinuity (intertidal) surface at Tynant, to a palaeokarst at Minera. The regression occurring on the Mesothem D5a / D5b boundary, therefore, was no more extensive than regressions between individual minor cycles.

The Potts Beck Limestone of the Ravenstonedale 'trough', of comparable lower Asbian (and Mesothem D5a) age (George et al 1976), to the Tynant and Whitehaven formations has a similar cycle pattern (section 5. 1. 2).

The Woo Dale Beds of North Derbyshire (Stevenson and Gaunt, 1971) allocated to the "S_2" (Holkerian) by the presence of Davidsonina carbonaria, but suggested as equivalent to early Asbian by George et al (1976, p.32), comprise a sequence of similar peri-tidal-subtidal cyclic sediments (interpreted from Stevenson and

Gaunt op cit, p. 37 - 43), that contain a faunal assemblage including Daviesiella sp, Dibunophyllum bourtonense; Lithostrotion martini and Linoprotonia hemisphaerica. Apart from the negative evidence offered by the lack of Davidsonina carbonaria in the North Wales succession, the Woo Dale Beds are biostratigraphic equivalents to the Whitehaven Formation ('early Asbian') of North Wales. Further parallels are readily drawn: Both bryozoan and oolitic facies are recorded in the Woo Dale Beds, comparable to developments in the Llanymynech Member. The equivalent Woo Dale Bed MA.1 suites (Lithofacies. L.5; calcilutites) are "very irregular and lenticular in character, swelling out and thinning rapidly" (Stevenson and Gaunt op. cit. p. 41).

The Chee Tor Rock, overlying the Woo Dale Beds and lateral equivalent to the Bee Low Limestone, has a cyclic style similar to the upper Asbian Eglwyseg and (basal part of the) Llynclys Formations.

12.2.1.2. MESOTHEM D5b. The transition from the cycle form of Mesothem D5a to D5b, in both the Llangollen and Oswestry embayments is gradational, with occasional peritidal facies within lower minor cycles of Mesothem D5b.

As with Mesothem D5a, basal argillaceous and thin bedded phases (lithofacies L.2; L.3 & L.6) of the Llynclys Formation are poorly developed. However, in higher cycles of this formation, the development of stylonodular sediments (Lithofacies L.4) intercalated with maroon shales reflects the regression and offlap associated with the Mesothem D5b / D6a boundary, with a significant input of argillaceous clastics from the Lower Palaeozoic hinterland.

Towards Minera the Eglwyseg Formation thins from its 'depocentre' at Tynant, and possibly has offlap associated with its upper cycles also. Although Somerville (1979a) correlated the upper cycle of the Eglwyseg Formation developed at Minera with his Minor Cycle 6 at Tynant, there is some doubt concerning this. Certainly, not all the upper cycles of the Eglwyseg Formation are fully developed at Minera. On the southern margin of the Llangollen embayment (at Froncysyllte) the situation is even more extreme, with just 5m of Mesothem D5b overlain by the transgressive facies of (the Brigantian) Mesothem D6a.

The cycle form of Mesothem D5b is also similar in sequences on separate depositional platforms and shelves in the North of England. (see chapter 5).

The Kingsdale Limestone (Mesothem D5b) of the Askrigg area rests with apparent non sequence upon Holkerian strata (Ramsbottom, 1974), and comprises an approximately equivalent number of minor cycles (Schwarzacher, 1958, chapter 5) to the Eglwyseg Formation. Variations between the cyclicity of the Askrigg and North Wales regions are accounted for by shelf palaeogeography and local tectonics primarily (see next section).

Individual cycles may not at present be correlated between these two areas.

12.2.1.3. <u>MESOTHEMS D6a and D6b.</u> In North Wales the base of the Brigantian appears to correspond to the base of Mesothem D6a, although according to Wedd <u>et al</u> (1927) and Wedd <u>et al</u> (1929) there is some palaeontological evidence to warrant inclusion of some of the uppermost Eglwyseg and Llynclys Formation sediments within the Brigantian on account of a ?Lonsdaleoid fauna. This has not been corroborated here.

The minor cycle style within the Trefor Formation is different from that of underlying mesothems in gross development, although individual minor cycles reflect similar palaeoenvironment transitions as observed in the Asbian strata. Dominated by thin-bedded argillaceous packstones and wackestone basal phases (Lithofacies L.3) these cycles represent transgression on a more widespread scale than minor cycles of underlying mesothems, as shown by their continuation across the Berwyns and extension at least as far west as Corwen (Morton 1879) . Although Morton (1879) reported "Middle White Limestone" at Corwen, this must now be in doubt in the light of recent work. No exposures are now visible apart from upper beds of the Trefor Formation, containing Orionastraea.

Ramsbottom (1977) split his original Mesothem D6 into D6a and D6b. The upper Brigantian is poorly exposed in the Llangollen and Oswestry areas (Sandy Passage Formation). The incoming of the Orionastraea - Corwenia fauna of the uppermost cycle exposed on Trefor Rocks (above Tf.) at the base of the Sandy Passage Formation, is visible as a ~20m thick thin-bedded and argillaceous transgressive phase, correlateable over the Llangollen embayment by both exposure and feature (chart D). This may represent the basal D6b transgression . Southwards within the Oswestry area correlation is tentative and this base of Mesothem D6b may be the base of the Sandy Limestone as suggested by Wedd et al (1929).

In the North of England equivalent Brigantian strata include the Yoredale cyclothems that have a characteristic basal dark thin-bedded argillaceous limestone, overlain by shoaling arenaceous fluviodeltaic progradational facies.

The Cefn Mawr Limestone Formation of the Vale of Clwyd, chronostratigraphic equivalent to the Trefor Formation, comprises a sequence of 8 minor cycles (Somerville, 1977, 1979c). However, in this thicker more distal shelf sequence regressive upper phases of cycles are well developed and facies are laterally more variable (e.g. development of crinoid 'bioherms', (Somerville, 1977)) than in more proximal shelf environments of Llangollen and Oswestry.

12.2.2. Lateral variation of cycle form.

12.2.2.1. EARLY ASBIAN MINOR CYCLES. Lateral variation in minor cycle form of the early Asbian contemporary Whitehaven and Tynant Formation is striking. Sychtyn Member minor cycles, unlike the Tynant Formation, lack transgressive basal phases and palaeokarsts are common, indicating prolonged subaerial emergence on minor cycle boundaries. It is concluded that the Sychtyn Member, in representing a more proximal shelf environment was deposited towards the transgression 'feather edge' against the Berwyns, where the water depth rarely exceeded fair-weather wave base, and sediment deposition lasted a shorter time period than corresponding minor cycles of the Tynant Formation. This is idealised in fig. 70, where the outcrop of the Sychtyn Member represents the shelf region towards "X". In contrast, the Tynant Formation, being ?more distal shelf and/or subject to greater subsidence rates also represents a position near "Y" on fig. 70. Minor cycles of the Tynant Formation also reflect more proximal shelf deposition towards the Cyrn y Brain.

Fig.77 models the depth of shelf flooding estimated for this cycle form showing that in more proximal shelf environments water depth at maximum transgression remains above fair weather wave base.

MODEL FOR MINOR CYCLE INTERRELATION

MESOTHEM D5A

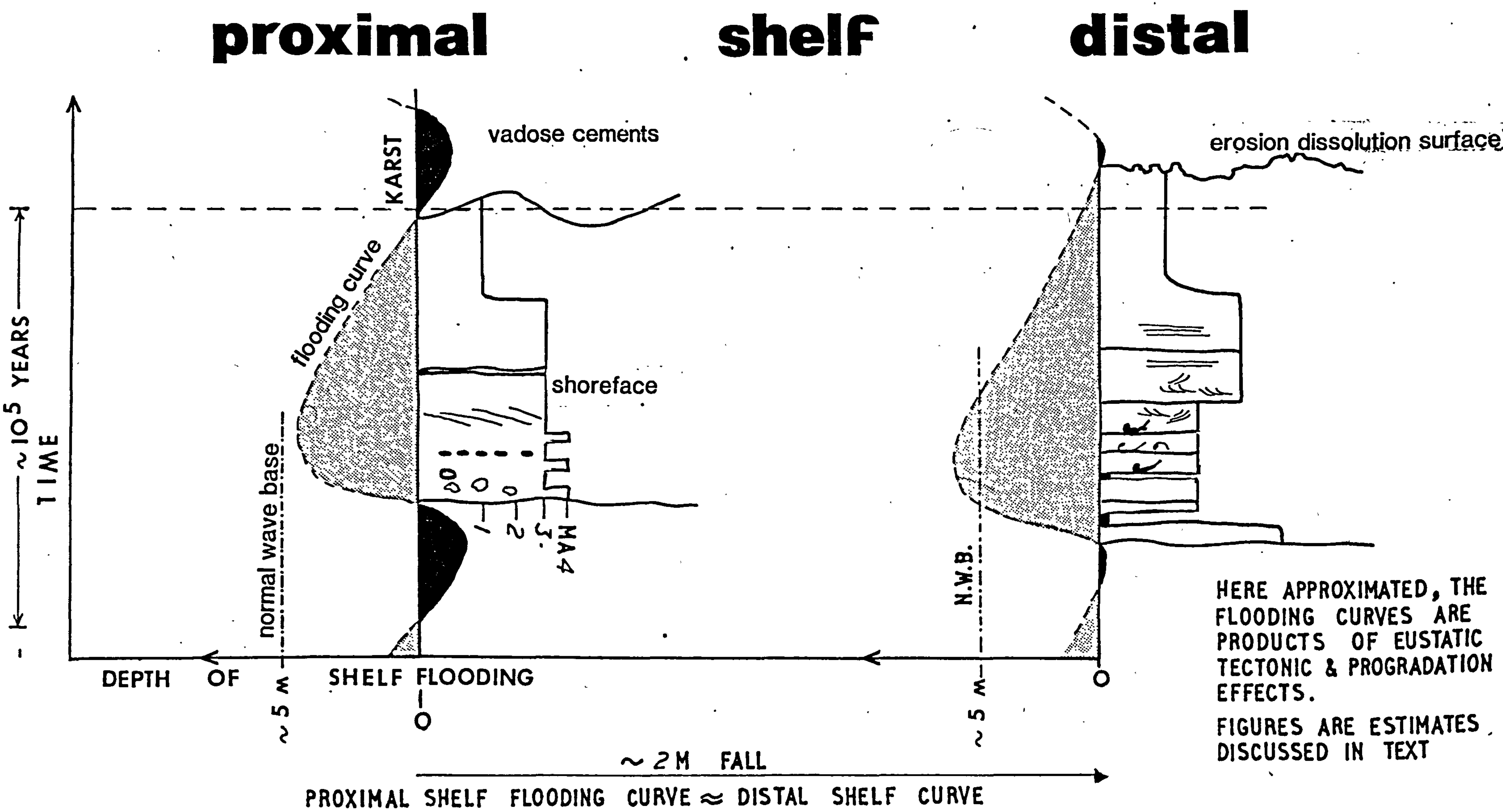


FIG 77

DEPOSITIONAL GRADIENTS. The development of peritidal facies lateral to palaeokarsts as regressive phenomena requires some explanation. Peritidal sedimentation occurs by progradation and is a regressive process. Superimposed rapid eustatic regression may affect this process. Depositional gradients play an important role in determining the rates and development of transgressions (and regressions) on shelves (Coogan 1969).

Recent carbonate environments have depositional gradients in the order of $\ll 1$ to 2m/km (e.g. Coogan, 1969; Enos and Perkins, 1977, p.136). The outcrop pattern in the Asbian of Llangollen and Oswestry is broadly "palaeocoast parallel" and, as such, depositional gradients were probably subdued across the outcrop, and would be most marked in areas of differential subsidence.

Few individual Tynant Formation minor cycles can be correlated a distance of about 7km between Tynant and Minera. Whilst basal sediments (MA.2) of minor cycles at Tynant were deposited below fair weather wave base but in waters ?5 - 10m, either subtidal (above wave base) or peritidal sediments were being deposited at Minera. A maximum depositional gradient for these sediments is, therefore, $<$ 2m/km, and probably $<$ 1m/km.

Within the Whitehaven Formation the transition from cycles of Sychtyn to Llanymynech Member forms appears to be relatively rapid (over about 1½km). The paucity of well-defined emergence horizons within the Llanymynech Member, and lack of peritidal facies suggests that prolonged emergence was rare and subtidal deposition may have continued across many 'cycle boundaries'. The chrono-facies model for regressive cycles (fig. 6 based on Coogan, 1969) applies. to the Llanymynech Member which represents a more distal shelf environment to the Sychtyn Member - readily acceptable from the palaeogeographic model of fig. 4.

12.2.2.2. UPPER ASBIAN MINOR CYCLES. The Eglwyseg Formation is as susceptible to palaeogeographic constraints as the underlying Lower Asbian. In the Minera area, where the Formation is 56m thick, basal bedded MA.2 argillaceous phases of minor cycles are absent, indicating transgression did not exceed water depths below fair weather wave base. Peritidal sediments are, however, common in the lower section (chart E) ?equivalent to minor cycle Eg.1.2 varying considerably laterally. ?Shoreface sediments also occur, towards regression (subaerial) surfaces (section 3.1.2.11) with extraformational conglomerates, rudstones and rudstone coquinas.

The situation is, however, complex, for these Minera peritidal MA.1 facies are not overlain by obvious palaeokarsts, and the rapid attenuation of the uppermost Tynant Formation regressive phase in the Minera area suggests local tectonics may have contributed to the sedimentation pattern.

In the most proximal environments of the latest Asbian in the Oswestry embayment, the influence of the St. George's Land hinterland on the offlapping minor cycles of Mesothem D5b is marked by a high content of clays, much reddened (diagenetic) by disseminated haematite.

The development of stylonodular (Lithofacies L.4) transgressive phase sediments above Eg.5; Eg.6; Eg.8 and within upper cycles of the Llynclys Formation appear to reflect relatively shallow transgressive events associated with minor cycles in offlap sequences.

12.2.2.3. BRIGANTIAN MINOR CYCLES. The lateral variation in minor cycles of the Trefor Formation reflects relative subsidence and extension of the transgressive seas. Chapter 4 showed a trend to reduced thickness of basal transgressive phases away from the depocentre of the Llangollen area, with regressive phase sediments accounting for a higher fraction of the sedimentary sequence in more proximal environments not subject to relative subsidence (fig.30).

The extensive nature of the D6a (Brigantian) transgression reflects low depositional gradients across the shelf and infill of both pre-existing and contemporaneously subsiding 'embayment' centres. The rapid thickening of the Trefor Formation into the more northerly Cefn Mawr Limestone Formation (Somerville 1979c) north of Minera reflects a transition from more proximal to distal shelf facies northwards from Oswestry, along the present north/south outcrop in part enhanced by the presence of the Bryneglwys Fault (fig. 3).

12.3. PROBLEMS ASSOCIATED WITH MINOR CYCLE CORRELATION.

12.3.1. Composite Cycles.

Schwarzacher (1975) defined 'composite' cycles as the product of more than one rhythmic mechanism reflected in the sedimentary sequence. He (op.cit.) recognised two types, defined by whether the two cyclic mechanisms act in a similar manner on the sedimentary record, or whether they produce a superimposed sequence of cycle types that are readily differentiated from each other.

The dominant cycle influences in the Lower Carboniferous of North Wales appear to have caused sea level fluctuations. These fluctuations produce type B composite cycles of Schwarzacher (1975).

The mesothemic/cyclothemic classification of Ramsbottom (1977, p.283) conforms to this composite cycle model. Recognition of the smallest significant cyclic effect in this sequence of diminishing wavelength cycles is important in determining the cyclic processes.

In the North Wales succession, 'minor cycles' are the most frequent cycle form. Theoretically, minor eustatic effects are more readily conveyed to the sedimentary record in more proximal-shelf environments, where only small fluctuations in water depth may produce significant changes in the palaeoenvironment and sedimentation style. Many of the 'multiple' cyclic boundaries observed in the Eglwyseg Formation (e.g. beneath Eg.9) may represent a small scale cyclic process producing sea level fluctuations that are not discerned in more distal shelf environments or during earlier transgressive phases of the minor cyclicity. The problem becomes one of degree: When does a sediment packet warrant full 'cycle' status, and when not?. This is not readily resolved as many cycles are laterally very variable. In the Tynant and Whitehaven Formation, since the cyclicity is on a small (commonly 50cm to 2m) scale, all minor cycles warrant equivalent terminology status, but even these cycles occasionally have multiple emergence events (see Chart C) separated by only a few centimetres of sediment. This highlights the problems involved in erection of strict cycle numbering schemes.

12.3.2. <u>Convergence and its Recognition.</u>

Traced to more proximal shelf positions, each minor cycle will become thinner as a consequence of reduced duration of marine sedimentation. This has been noted by Enos and Perkins (1977,p.183) on the Pleistocence shelf of Florida as 'up-dip convergence of subaerial emergence surfaces'. Such convergence leads to compound emergence events (fig.70).

Compound emergence features may be formed either by the alteration and / or removal of a thin carbonate unit sandwiched between emergence horizons, or by continued subaerial emergence of a proximal shelf position during offlap of successive cycle(s). This situation explains, on a large scale, the Tynant / Eglwyseg Formations' junction at Froncysyllte, and probably occurred during offlap of upper cycles of the Llynclys and Eglwyseg Formations. Significantly, in both the Minera and Froncysyllte succession there is no evidence of compound emergence events towards the end of the Asbian (Mesothem D5b). Indeed, the general lateral variability of many emergence horizons promotes speculation as to whether relative duration of individual emergence events can be estimated. Walkden (in Somerville, 1979a, p.338) noted that the 'degree of development' of emergence phenomena may not be a crucial factor in establishing correlations.

The Tynant and Whitehaven Formations represent most proximal shelf situations, and it is, therefore, within these deposits that minor compound emergence events may have occurred. Their recognition, however, remains questionable due to the lateral variability of cycle form, and poor outcrop. Indeed, their presence depends greatly upon the gradient across the shelf. A shelf with negligible gradient will very rarely preserve compound emergence events except in extreme proximal environments.

Rare examples of minor compound emergence events occur in the Oswestry and Llangollen areas as just one emergence surface that is lateral equivalent to a multiple emergence sequence. Probable example include Tf.3 (Trefor Formation, fig. 30), and another within the Llanymynech Member at Llanymynech (chart C). Both have planar erosion surfaces / palaeokarsts overlying a S.D. surface that is laterally equivalent to a palaeokarst / planar erosion surface. It is the lack of recognition of planar erosion surfaces (cf. Read and Grover 1977) within homogeneous regressive phases that hinders the recognition of more compound emergence surfaces within the Whitehaven and Tynant Formations.

Part E

CONCLUSIONS

13. CONCLUSIONS

The Lower Carboniferous of North Wales, in the vicinity of Llangollen and Oswestry, records the deposition of cyclic shallow marine shelf carbonates during a gradual transgression on to the north eastern flanks of the Anglo-Welsh Landmass (St. George's Land).

This sequence of carbonates spans the major eustatic cycles (Mesothems) D5a to the base of D6b of Ramsbottom (1977), approximately equivalent to the Asbian and Lower Brigantian of George (*et al* 1976).

Lower Palaeozoic structural elements of St. George's Land influenced both local palaeogeography and sedimentation.

Throughout the Asbian, the Berwyn Hills provided a positive E. W trending peninsula that remained subaerially emergent until the lower Brigantian dividing the study area into two depositional embayments; a northerly one centred near Llangollen, and a southerly one centred near Oswestry.

The Cyrn y Brain provided a similar feature on the Northern flanks of the Llangollen embayment, whilst the eastward extension of St. George's Land itself provided a southerly limit to the Oswestry embayment, and transgressing Lower Carboniferous seas.

The Cyrn y Brain was overstepped, along present outcrop, by early Asbian sediments, although it remained a positive influence into the Brigantian.

Synsedimentary local subsidence perpetuated the basement influences.

A new lithostratigraphy is proposed for the Oswestry area, comprising:

Mesothem D5b (Upper Asbian): Llynclys Formation

Mesothem D5a (Lower Asbian): Whitehaven Formation, divided into the Sychtyn and Llanymynech Members.

The recognition of 14 microfacies (based upon petrographic criteria), grouped into 4 microfacies associations, and 8 lithofacies (based upon field criteria) have provided the framework for the petrographic analysis and palaeoenvironment interpretation. Tables 1 and 2 list these microfacies and lithofacies respectively.

13.1. GENERAL STRATIGRAPHIC CONCLUSIONS

13.1.1. Early Asbian sediments of Mesothem D5a.

At least 20 minor cycles were deposited in the Llangollen embayment during the early Asbian (Mesothem D5a), each reflecting proximal shelf environments with progradation of peritidal facies. over transgressive phase subtidal facies.

Within the Oswestry embayment, seaward (and southward) of similar peritidal-regressive minor cycles, a shoaling environment restricted progradation of peritidal facies, with the formation of oolitic sand blankets and cross-bedded units, in a sequence lacking many well-defined minor cycle boundaries

Dissimilarities in the peritidal regressive minor cycles of both embayments reflect palaeogeography and palaeotectonics. The Whitehaven Formation (Oswestry) succession, and Tynant Formation (Llangollen succession) deposited upon the axis and flanks of the Cyrn y Brain, represent more proximal shelf environments less subject to relative subsidence than the Llangollen embayment depocentre. These proximal areas suffered more prolonged subaerial emergence on regressions. Basal argillaceous transgressive phases of the minor cycles are minor or absent in these proximal areas, reflecting water depth variations and sedimentation above normal wave base in more proximal areas during the transgressive phase in comparison to more distal environments.

These minor cycles are mostly between 1 and 4m thick.

The regressive peritidal facies indicate penecontemporary emergence and have a suite of shallow subtidal to supratidalindicators, including vadose cements, penecontemporary dolomites, peds, and ?rootlet horizons, but lack evaporites.

Comparison of fenestrae distributions and forms with Recent environments indicates that .the non-random cyclicity in fabrics from tubular to irregular to laminoid forms is consistent with a progradational model.

13.1.2. Late Asbian Sediments of Mesothem D5b.

The transition to sedimentation of late Asbian Mesothem D5b was, in part, palaeogeographically / palaeotectonically controlled.

Sedimentation was virtually continuous across the mesothem boundary within the Llangollen embayment depocentre, but subaerial emergence was marked elsewhere.

At its maximum development at least fourteen minor cycles developed in the Llangollen embayment.

These minor cycles are dominated by an upper massive bedded alga peloid grainstone phase, often pseudobreccia mottled. Regressive phases of calcisphere wackestone, unlike Mesothem D5a, are mostly absent. Minor cycle tops are marked by calcretes, palaeokarsts and palaeosols. At times these regression events are multiple attesting to minor rhythmicity in strand-line regressive trends.

Transgressive phases of these minor cycles are palaeotopographically controlled in lateral extent, thinning towards tectonically positive features in a similar manner to transgressive phases of Mesothem D5a minor cycles.

Regression and subaerial emergence appears to have been rapid as indicated by the lack of progradational peritidal facies. This may reflect a shallow depositional gradient across the shelf.

Subaerial emergence phenomena are variable. Calcrete structures are preserved upon surfaces with little palaeokarstic relief, but are lacking on deeply karstified surfaces. Calcretes predate karstification.

Towards the end of Mesothem D5b cycle offlap occurred in the Oswestry area, and may account for the attenuated sequence over the Cyrn y Brain at Minera. In the Oswestry area, this offlap was marked (upper cycles of the Llynclys Formation) by the development of thin cycles ($\leq$ 5m) with haematitic and argillaceous-stylonodular basal phases, suggesting derivation of terrigeneous clastics from proximal and weathered terrains.

13.1.3. Mesothem D6a Trefor Formation ('Tf.1 - Tf. 8')

The basal Brigantian transgression in the Llangollen and Oswestry areas produced an initial sequence up to 20m of thin bedded argillaceous packstones and wackestones. Thickest over the Llangollen embayment (70m), this Formation transgressed over the Berwyn peninsula, but thins southward to ~ 25m at Dolgoch, and includes at least 7 minor cycles

Most of this thinning is a consequence of reduction in development of argillaceous transgressive phases of minor cycles into more proximal shelf environments.

Laterally persistent biostromes pervade the sequence, and are especially prominent in lower thin bedded argillaceous phases.

Tectonic activity is indicated by both differential subsidence, and loaded horizons suggest penecontemporary seismic activity.

The upper boundary is redefined along Tf.8 in this study.

13.1.4. Mesothem D6b.

The transgressive phase, incorporating 'Morton's Coral Bed' may represent the base of Mesothem D6b of Ramsbottom.

This transgressive event marks the base of the Sandy Passage Formation, as defined herein, and removed from the Trefor Formation.

Minor bioherms within the basal transgressive phase of Mesothem D6b attest to growth-baffle formation, through initiation around large stable colonies of cerioid lonsdaleoids and lithostrotiontids that, in part, entrapped less stable locally transported coral colonies.

13.2. EMERGENCE AND PALAEOCLIMATE.

Large areas of the carbonate shelf were subaerially exposed during regression at minor cycle boundaries. This emergence is characterised by modification of the carbonates by subaerial agencies, primarily the development of calcrete features, palaeokarsts and palaeosols.

1. When preserved, calcretisation of emergence surfaces occurred

prior to palaeokarstic dissolution. Both processes took place individually over a period of the order of $\sim 10^4$ years but were not synchronous. They may reflect a palaeoclimatic cyclicity, between arid and pluvial phases, analogous to Pleistocene palaeoclimatic pluvial cycles.

2. A lateral variability of emergence profiles, in part dependent upon palaeogeographic/tectonic constraints, with more proximal shelf areas reflecting most prolonged emergence, whilst in distal areas strand-line features are preserved, including 'sutured discontinuity surfaces'.

3. Sutured discontinuity surfaces are irregular centimetre relief bedtop morphologies that are primary features formed by chemical biological and mechanical agencies within an intertidal environment. They formed during minor cycle regression, and attest to the maximum limit of the regression.

They are laterally equivalent to planar erosion surfaces and palaeokarsts, and probably formed over a period of a few thousand years.

13.3. CYCLE MECHANISMS.

Minor cycle form was controlled in part by eustatic sea level fluctuations, and in part by sediment progradation.

Lateral variation in cycle form reflects palaeotopography that was controlled by active tectonism. Minor cycles were thickest throughout both the Asbian and Brigantian in areas of active subsidence. This thickening is most apparent within basal thin bedded and argillaceous 'transgressive' phases of cycles, which are absent or reduced on more proximal shelf and relatively tectonic positive positions. This feature is also apparent in other Asbian carbonates of the Askrigg area. Whether subsidence was a continuous or spasmodic process is unclear.

13.4. MICROFACIES ANALYSIS.

Microfacies analysis of the sediments has led to the recognition of four microfacies association, each comprising an assemblage of field- related microfacies (approximately equivalent to lithofacies).

Microfacies associations recognised are:

Calcisphere wackestone (MA.1)

Argillaceous alga packstone (MA.2)

Alga peloid grainstone (MA.3)

Bioclast peloid rudstone (MA.4)

MA.1 sediments are characteristic regressive phases of Mesothem D5a minor cycles, with a varied suite of intertidal and supratidal phenomena. They are minor sediments within regressive phases of both upper Asbian and Brigantian (Mesothems D5b and D6a) minor cycles.

MA.2 microfacies characterise transgressive phases of minor cycles that are dominated in the Asbian, by the tubular-septate beresellid Kamaenella denbighi. These sediments contain a diverse macrofaunal assemblage. They reflect a subtidal environment, below a normal (fair weather) wave base, and indicate the assymetry of minor cycles and probable rapidity of minor cycle transgressions.

MA.3 microfacies characterise regressive phases of minor cycles, and were apparently deposited above or near to, a normal wave base. The clasts are very variable and indicate sediment movement and admixing by their mixed generation form. Storm deposits occur primarily within units dominated by MA.3 microfacies, as bioclast rudstones, brachiopod coquinas, and layers of fragmentary macrofossils.

Within Brigantian sediments, the dasycladacean Coelosporella replaces Koninckopora and beresellid algae that were two important sediment contributors in the Asbian.

Ternary plots of bioclast compositions, plotted as calcispheres + gastropods (C); algae (A): echinoderm + brachiopod + coral + bryozoan (M) show basic trends within the bioclast components of minor cycles, from sediments with high algal content in minor cycle basal phases to a more mixed assemblage of bioclasts and finally into peritidal regressive facies that plot towards the 'C' end member.

Standardised petrographic diversities provide data for assessing the degree of palaeoenvironment restriction by comparison between penecontemporary microfacies associations. Sediments of peritidal suite (MA.1 microfacies) have ~~pt~~petrographic diversities about half that of subtidal suite sediments.

Whilst it is accepted that bioturbation may initiate much of the pseudobreccia mottling observed within Asbian and Brigantian MA.3 microfacies, differential cementation (early diagenetic) is the most important mottle forming process.

Intermottle matrix (poorly cemented) is preferentially compacted and microstylolitised to intramottle sediment, and in extreme cases, preferential loss of primary aragonite moulds within intermottle sediment produces pseudointraclastic textures (i.e.,Brigantian MA.3 mottled sediments).

REFERENCES

ADELEYE, D.R., 1973. Origin of ironstones, an example from the middle Niger Valley, Nigeria. J. sedim. Petrol., 43, 709 - 727.

AGER, D.V., 1963. Principles of paleoecology. Mc.Graw-Hill, New York. 371pp.

AIGNER, T., HAGDORN, H., and MUNDLOS, R., 1979. Biohermal, biostromal and storm-generated coquinas in the Upper Muschelkalk. Neues Jb. Geol. Paläont. Abh., 157, 42-52.

ALLEN, J.R.L., and BANKS, N.L., 1972. An interpretation and analysis of recumbent-folded deformed cross-bedding. Sedimentology, 19, 257 - 283.

ALLING, H.L., 1947. Diagenesis of the Clinton Haematite ores of New York. Geol. Soc. Am. Bull., 83, 59 - 86.

ASSERETO, R.L.A.M., and KENDALL, C.G. St.C., 1977. Nature, origin and classification of peritidal tepee structures and related breccias. Sedimentology, 24, 153 - 210.

BADIOZAMANI, K., 1973. The Dorag dolomitization model-application to the Middle Ordovician of Wisconsin. J. sedim. Petrol., 43, 965- 984.

———, MACKENZIE, F.T., THORSTENSON, D.C., 1977. Experimental carbonate cementation: salinity, temperature and vadose-phreatic effects. J. sedim. Petrol., 47, 529 - 542.

BALL, M.M., 1967. Carbonate sand bodies of Florida and the Bahamas. J. sedim. Petrol., 37, 556-591.

———, SHINN, E.A. and STOCKMAN, K.W., 1967. The geologic effects of hurricane Donna in south Florida. J. Geol., 75, 583 -597.

BALL, S.M., 1971. The Westphalia Limestone of the northern mid-continent: a possible ancient storm deposit, J. sedim. Petrol., 41, 217 - 232.

BANERJEE, A., 1969. Carboniferous lithofacies Rhydymwyn to Erryrys, North Wales, Geol. J., 6, 181-184.

BATHURST, R.G.C., 1966. Boring algae, micrite envelopes and lithification of molluscan biosparites. Geol. J., 5, 15-32.

———, 1975. Carbonate sediments and their diagenesis, 2nd ed., Elsevier, Amsterdam, 658pp.

BATTEN, R.L., 1966. The Lower Carboniferous gastropod fauna from the Hotwells Limestone of Compton Martin, Somerset. Part 1. Palaeontogr. Soc. [Monogr.], 1, 52pp.

BAUGHEN, D.J., and WALSH, P.T., 1980. Palaeokarst phenomena in the Carboniferous Limestone of Anglesey, North Wales. Trans. British Cave Research Assoc., 7, 13-30.

BEHRENS, E.W., and LAND, L.S., 1970. Subtidal Holocene dolomite, Baffin Bay, Texas. J. sedim Petrol., 42, 155 - 161.

BEMROSE, H.H.A., 1907. The toadstones of Derbyshire; their field relations and petrography. Q.Jl. geol. Soc. Lond . 63, 241 - 281.

BERNER, R.A., 1969. Goethite stability and the origin of red beds. Geochim. Cosmochim. Acta, 33, 267-273.

BLACK, M., 1933. The algal sediments of Andros Island, Bahamas. Phil. Trans. R . Soc. London (B), 222, 165-192.

BONTE, A., 1963. Les remplissages karstiques. Sedimentology, 2, 333-340.

BOSE, P.K., 1979. Penecontemporaneous dolomitization in the Precambrian Bhander Limestone, Rajasthan, India - a petrographic attestation. Geol. Rund., 68, 680-695.

BOTT, M.H.P. and JOHNSON, G.A.L., 1967. The controlling mechanism of Carboniferous cyclic sedimentation. Q. Jl. geol. Soc. Lond., 122, 421- 441.

BOURROUILH, F, 1978. Bahamian Tidal Flats Hydrology, NW Andros. First Results. Relationships with dolomitisation.(Abstrac In Titles and Abstracts of talks. Sixth meeting of carbonate sedimentologists, Dept of Geology,Univ.Liverpool

BOWEN, D.Q., 1978. Quaternary geology. A stratigraphic framework for multidisciplinary work. Pergamon, Oxford.

BOWMAN, M.B.J., 1979. The depositional environments of a limestone unit from the San Emiliano Formation (Namurian/Westphalian) Cantabrian Mts., N.W. Spain. Sedim. Geol., 24, 25-43.

BOUCOT, A.J. and GRAY, J., 1979. Epilogue: A palaeozoic Pangaea. In GRAY, J., and BOUCOT, A.J. (eds). Historical biogeography, plate tectonics and the changing environment. Corvallis, Oregon State University, 465-482.

BRADSHAW, M.J., JAMES, S.J., and TURNER, P., 1980. Origin of oolitic ironstones - discussion. J.sedim. Petrol., 50, 295-304.

BRAITHWAITE, C.J.R., 1975. Petrology of palaeosols and other terrestrial sediments on Aldabra, Western Indian Ocean. Phil. Trans. R. Soc., B273, 1-31.

BRENCHLEY, P.J., NEWALL, G., and STANISTREET, I.G., 1979. A storm surge origin for sandstone beds in an epicontinental platform sequence, Ordovician, Norway. Sedim. Geol., 22, 185-217.

BREWER, R., 1964. Fabric and Mineral Analysis of soils. Wiley, New York, 470 pp.

BROMLEY, R.G., 1968. Burrows and borings in hardgrounds. Meddr. dansk geol. Foren., 18, 247-250.

———, 1975. Trace fossils at omission surfaces. In FREY. R.W. (ed). The study of trace fossils. Springer-Verlag, New York, 399-428.

BURGESS, C.J., 1978. Holocene carbonate sedimentation Lakeba and Vanua Balavu, Lau, Fiji, Bull. R . Soc. New Zealand, 17, 191-209.

BUSHINSKIY, G.I., 1961. Stylolites. Izv. Akad. Nauk. SSSR, Ser. Geol., 8, 31-46 (English translation)

CASSHYAP, S.M., 1975. Cyclic characteristics of coal-bearing sediments in the Bochumer Formation (Westphal A.2) Ruhrgebiet, Germany. Sedimentology, 22, 237 - 255

CAYEAUX, L., 1935. Les Roches sedimentaires de France: roches carbonates. Masson, Paris. 447pp.

CLOUD, P.E., 1960. Gas as a sedimentary and diagenetic agent. Am. J. Sci. 258, 35-45.

CLOUD, P.E., 1962. Environment of calcium carbonate deposition west of Andros Island, Bahamas, U.S. Geol. Surv. Prof. Paper. 350, 138pp.

CONIL, R., and LYS, M., 1964. Materiaux pour l'etude micropaleontologique du Dinantien de la Belgique et de la France (Avesnois). Mem. Univ. Louvain, Inst. Geol., 23, 1 - 296.

COOGAN, A.H., 1969. Recent and ancient carbonate cyclic sequences. In ELAM, J.C., and CHUBER, S. (eds). Cyclic sedimentation in the Permian Basin, 2nd ed., 5-16, Midland Texas, West Texas geol. Soc., 1972.

COPE, F.W., 1939. The mid-Visean (S_2 - D_1) succession in north Derbyshire and north west England. Proc. Yorks. Geol. Soc., 24, 60-66.

CORBEL , J., 1959. La vitesse de l'erosion. Zeit. Geomorph., 3, 1 -28.

CORRENS, C.W., 1950. Zur geochemie der diagenese. 1. Das verhalten von $CaCO_3$ und SiO_2. Geochim.Cosmochim.Acta, 1, 49-54.

COULTER H.W., MIGLIACCIO, R.R., 1966. Effects of the earthquake of March 27, 1964 at Valdez, Alaska. U.S. Geol. Surv. Prof. Paper. 542C, 36pp.

DAETWYLER, C.C., and KIDWELL, A.L., 1959. The Gulf of Batabano, a modern carbonate basin. World Petrol. Congr. Proc., 5th New York, 1959, Section 1, paper 1, 1 -21.

DAVIDSON, T., 1874 - 82. A monograph of the British Fossil brachiopoda, 4. Palaeontogr. Soc. [Monogr]. , London, 383 pp.

DAVIES, D.C., 1865. The outlier of Carboniferous Limestone near Corwen, North Wales. Geol. Mag., 2, 283-285.

DAVIES, J.L., 1964. A morpho genic approach to world shorelines. Zeit. Geomorph., 8, (Spec. No.) 127-142.

DEELMAN, J.C., 1972. On mechanisms causing birdseye structures. N. Jb. Geol. Paläont. Mh., 1972, 582-595.

DEFFEYES, K. S., LUCIA. F.J., and WEYL, P.K., 1965. Dolomitization of Recent and Plio-Pleistocence sediments by marine evaporite waters on Bonaire, Netherlands, Antilles. Soc. Econ. Pal-eontologists Mineralogists Spec. Publ. 13, 71 - 88.

DIECI, G., RUSSO, A., RUSSO, F., and MARCHI, M.S., 1977. Occurrence of spicules in Triassic chaetetids and ceratoporellids. Boll. Soc. paleont. ital., 16, 229-238.

DIXON, E.E.L., 1909. Limestone unconformities and their contemporaneous pipes and swallow holes. Geol. Mag., 6, 514 - 516.

DONOVAN, D.T., and JONES, E.J.W., 1979. Causes of world-wide changes in sea-level. Q.Jl. geol. Soc. Lond., 136, 187-192.

DRAVIS J., 1979. Rapid and widespread generation of Recent oolitic hardgrounds on a high energy Bahamian Platform, Eleuthera Bank, Bahamas. J. sedim.Petrol., 49, 195-208

DUFF, P.M.D., and WALTON, E.K., 1962. Statistical basis for cyclothems: a quantitative study of the sedimentary succession in the East Pennine Coalfield, Sedimentology, 1, 235-255

———, HALLAM, A., and WALTON, E.K., 1967. Cyclic sedimentation (Developments in sedimentology, 10) 280pp. Elsevier, Amsterdam.

DUNHAM, R.J., 1962. Classification of carbonate rocks according to depositional texture. In HAM, W.E. (ed.) Classification of carbonate rocks Am. Ass. petrol. Geol. Mem., 1, 108-121.

———, R.J., 1969a Early vadose silt in Townsend mound (reef), New Mexico. In FRIEDMAN, G.M. (ed). Depositional environments in carbonate rocks. Soc. Econ Paleontologists Mineralogists Spec.Publ. 14, 137-181.

DUNHAM, R.J., 1969b. Vadose pisolite in the Capitan Reef (Permian), New Mexico and Texas. In FRIEDMAN, G.M. (ed) Depositional environments in carbonate rocks. Soc. Econ. Paleontologists Mineralogists, Spec. Publ., 14, 182-191.

———, R.J., 1971. Meniscus cements. In BRICKER, O.P. (ed), Carbonate Cement. Johns Hopkins, Baltimore, 297-300.

ELLIOT, G.F., 1970. Calcareous algae new to the British Carboniferous. Palaeontology, 13, 443-450.

EMBRY, A.F., and KLOVAN, J.E., 1971. A late Devonian reef tract on northeastern Banks Island, N.W.T. Bull Can. petrol. Geol. 19, 730-781.

ENGH, L., 1980. Can we determine solutional erosion by a simple formula?; Trans. British Cave Research Assoc., 7, 30-32.

ENOS, P., and PERKINS, R.D., 1977. Quaternary sedimentation in south Florida. Geol. Soc. Am. Memoir, 147, 198pp.

ESTEBAN, M., 1976. Vadose pisolite and caliche. Am. Ass. petrol. Geologists, Bull. 60, 2048 - 2057

FAIRBRIDGE, R.W., 1954. Stratigraphic correlation by microfacies. Am. J. Sci., 252, 683-694.

FERGUSON, J., 1978. Some aspects of the ecology and growth of the Carboniferous gigantoproductids. Proc. Yorks. geol. Soc., 42, 41-54.

FISCHER, A.G., 1964. The Lofer cyclothems of the Alpine Triassic. Geol. Surv. Kansas, Bull. 169, 107 - 149.

———, A.G., 1975. Tidal deposits, Dachstein Limestone of the North Alpine Triassic. In GINSBURG, R.N. (ed) Tidal Deposits, Springer-Verlag, New York, 235-242.

FITZPATRICK, E.A., 1980. Soils. Their formation, classification and distribution. Longman, London, 353 pp.

FLUGEL, E., 1978. Mikrofazielle Untersuchungs - Methoden von Kalken. Springer -Verlag, Berlin. 454pp.

FOLK, R.L., and LAND, L.S., 1975. Mg/Ca ratio and salinity: two controls over crystallization of dolomite. Am. Ass. petrol. Geologists, Bull, 59, 60 - 68

FRAKES, L.A., 1979. Climates throughout geologic time. Elsevier, Amsterdam,

________, L.A., and CROWELL, J.C., 1969. Late Paleozoic glaciation:l South America. Geol. Soc. Am. Bull., 80, 1007-1042.

FREEMAN, T., 1964. Algal limestones of the Marble Falls Formation Lower Pennsylvanian) central Texas. Geol. Soc. Am. Bull., 75, 669-676.

FURSICH, F.T., 1979. Genesis, environments, and ecology of Jurassic Hardgrounds. N. Jb. Geol. Palaont. Abh., 158, 1-63.

GARETT, P., 1977. Biological communities and their sedimentary record. In HARDIE, L.A. (ed) Sedimentation on the modern carbonate tidal flats of Northwest Andros Island, Bahamas. Johns Hopkins University, Baltimore, 202pp

GARWOOD, E.J., 1913. The Lower Carboniferous succession in the north-west of England. Q. Jl. geol. Soc. Lond. 68, 449-596.

GARWOOD, E.J., and GOODYEAR, E., 1924. The Lower Carboniferous succession in the Settle district along the line of the Craven Fault. Q.Jl. geol. Soc.Lond., 80, 184-272.

GEBELEIN,C.D. 1978.Stratigraphy and diagenesis of Holocene carbonate muds , southwest Andros Island, Bahamas(Abstract) In Titles and abstracts of talks. Sixth Meeting of carbonate sedimentologists, Department of Geology,Univ. Liverpool, 3-7 January,1978.

GEBELEIN, C.D., STEINEN, R.P., HOFFMAN, E.J., GARRETT, P., and QUEEN, J.M., 1979. Mixing zone dolomite in tidal flat sediments of central-west Andros Island, Bahamas, Am. Ass . Petroleum Geologists,Bull 63, 457 (Abstract)

GEORGE, T.N., 1958. Lower Carboniferous palaeogeography of the British Isles. Proc. Yorks. geol. Soc., 31,227-318.

———, 1978. Eustasy and tectonics. Sedimentary rhythms and stratigraphic evidence in British Dinantian correlation. Proc. Yorks. geol. Soc., 42, 229-262.

———, JOHNSON, G.A.L., MITCHELL, M., PRENTICE, J.E., RAMSBOTTOM, W.H.C., SEVASTOPULO, G.D., and WILSON, R.B., 1976. A correlation of Dinantian rocks in the British Isles. Geol. Soc. Lond., Spec. Rep. No. 7, 87pp.

GERMANN, K., 1969. Reworked dolomite crusts in the Wettersteinkalk (Ladinian, Alpine Triassic) as indicators of early supratidal dolomitization and lithification. Sedimentology, 12, 257-277.

GINGERICH, P.D., 1969. Markov analysis of cyclic alluvial sediments. J. sedim. Petrol. 39, 330-332.

GINSBURG, R.N., 1975. Recent Carbonate examples In GINSBURG R.N. (ed) Tidal deposits, Springer-Verlag , New York, 198-199

GINSBURG, R.N., and HARDIE, L.A., 1975. Tidal and storm deposits, northwestern Andros Island, Bahamas. In GINSBURG, R.N. (ed) Tidal deposits, Springer-Verlag, New York,201-208.

* Greensmith, J.T., Hatch, F.H and Rastall, R.H., 1971. Petrology of the sedimentary rocks. Murby, London, 502 pp. (Fifth edition)

GOUDIE, A., 1973. Duricrusts in tropical and subtropical landscapes. Clarendon, Oxford, 174pp.

GRAF, D.L., and GOLDSMITH, J.R., 1956. Some hydrothermal syntheses of dolomite and protodolomite. J. Geol., 64, 173-186.

GRAY, D.I., 1980. Spicule pseudomorphs in a new Palaeozoic chaetetid, and its sclerosponge significance. Palaeontology, 23, 803-820.

GREENLY, E., 1900. On sandstone pipes in the Carboniferous Limestone at Dwlban Point, East Anglesey. Geol. Mag. 4, 20-24.

* GROVER. Jr., G. and READ, J.F., 1978. Fenestral and associated vadose diagenetic fabrics of tidal flat carbonates, Middle Ordovician, New Market Limestone, southwestern Virginia. J. sedim.Petrol. 48, 453-473.

HAGAN, G.M. and LOGAN, B.W., 1974. History of Hutchinson embayment tidal flat, Shark Bay, Western Australia. In LOGAN,B.W., (ed) Evolution and diagenesis of Quaternary carbonate sequences, Shark Bay, Western Australia. Am. Ass. Petrol. Geologists Mem., 22, 283-315.

HAGAN, G.M. and LOGAN, B.W. 1975 Prograding Tidal Flat Sequences: Hutchinson Embayment, Shark Bay, Western Australia. In GINSBERG, R.N. (Ed) Tidal Deposits, Springer-Verlag, New York, 215 - 222.

HAM, W.E., 1952. Algal origin of the 'birdseye' limestone in the Molish Formation. Proc. Okla. Acad. Sci. 33, 200-203.

HANDFORD, C.R., 1978. Monteagle Limestone (Upper Mississippian) oolitic tidal-bar sedimentation in southern Cumberland Plateau. Am. Assoc. petrol. Geologists, Bull. 62, 644-656.

HANOR, J.S., 1978. Precipitation of beachrock cements: mixing of marine and meteoric waters vs. CO_2 degassing: J. sedim. Petrol., 48, 489-501.

HARDIE, L.A., 1977 (ed) Sedimentation on the modern carbonate tidal flat of Northwest Andros Island, Bahamas. Johns Hopkins, Baltimore, 202pp.

________, and GARETT, P., 1977. General environmental setting. In HARDIE, L.A. (ed). Sedimentation on the modern carbonate tidal flats of Northwest Andros Island, Bahamas, Johns Hopkins, Baltimore, 202 pp.

________, and GINSBURG, R.N., 1977. Layering: the origin and environmental significance of lamination and thin bedding. In HARDIE, L.A., (ed) Sedimentation on the modern carbonate tidal flat of Northwest Andros Island, Bahamas, Johns Hopkins, Baltimore, 202pp.

HARTMAN, W.D. and GOREAU, T.F., 1970. Jamaican coralline sponges: their morphology, ecology and fossil relatives. Symp zool. Soc. Lond., 25, 205-243.

________ ________, 1972. Ceratoporella (Porifera: Sclerospongiae) and the chaetetid corals. Trans. Conn. Acad. Arts Sci. 44, 133-148.

HAVARD, C., and OLDERSHAW, A., 1976. Early diagenesis in back-reef sedimentary cycles, Snipe Lake Reef Complex, Alberta. Bull. Can. petrol. Geology, 24, 27-69

HAY, R. L., and REEDER, R.J., 1978. Calcretes of Olduvai Gorge and the Ndolanya Beds of northern Tanzania. Sedimentology, 25, 649 - 673.

HEMINGWAY, J.E., 1973. Jurassic. In RAYNER, D.H. and HEMINGWAY, J.E., (eds) The geology and mineral resources of Yorkshire. Yorks. Geol. Soc. Occasional Publ.2, Leeds, 161-224

HEY, R.W., 1956. Cherts and limestones from the Crow series near Richmond, Yorkshire. Proc. Yorks geol. Soc., 30, 289-299.

HIND, W., and STOBBS, J.T., 1906. The Carboniferous succession below the Coal Measure in north Shropshire, Denbighshire and Flintshire, Geol. Mag. 3, 385-400, 445-459, 496-507.

HOFFMEISTER, J.E., and MULTER, H.G., 1965. Fossil mangrove reef of Key Biscane, Florida. Geol. Soc. Am. Bull., 76, 845-852.

HOLLAND, C.H., AUDLEY-CHARLES, M.G., BASSETT, M.G., COWIE, J.W., CURRY, D., FITCH, F.J., HANCOCK, J.M., HOUSE, M.R., INGHAM, J.K., KENT, P.E., MORTON, N., RAMSBOTTOM, W.H.C., RAWSON, P.F., SMITH, D.B., STUBBLEFIELD, C.J., TORRENS, H.S., WALLACE, P., and WOODLAND, A.W., 1978. A guide to stratigraphical procedure. Geol. Soc. Lond. Spec. Rep. 10., 18pp.

HOLMES, A., 1965. Principles of physical geology, Nelson, London, 1288pp

HOUBOLT, J.J.H.C., 1957. Surface sediments of the Persian Gulf near the Quatar Peninsula. Thesis, Univ. Utrecht, Utrecht, 113pp.

HSU, K.J., and SIEGENTHALER, C., 1969. Preliminary experiments on hydrodynamic movement induced by evaporation, and their bearing on the dolomite problem. Sedimentology, 12, 11-25.

HSU, K.J. and SIEGENTHALER, C., 1971. Preliminary experiments on hydrology of supratidal dolomitization and cementation. _In_ BRICKER, O.P. (ed), _Carbonate Cements._ Johns Hopkins, Baltimore, 315-318.

HUBBARD, J.A.E.B., 1970. Sedimentological factors affecting the distribution and growth of Visean caninoid corals in N.W. Ireland. _Palaeontology._ 13, 191-209.

HUDSON, J.D., 1962. Pseudopleochroic calcite in recrystallized shell-limestones., _Geol. Mag._ 92, 492 - 500.

HUDSON, R.G.S., and MITCHELL, G.H., 1937. The Carboniferous Geology of the Skipton Anticline. _Summ. Prog. Geol. Surv._ for 1935, (2) pp. 1 - 45

HUGHES CLARKE, M.W., and KEIJ, A.J., 1973. Organisms as producers of carbonate sediment and indicators of environment in the southern Persian Gulf. _In_ PURSER, B.H. (ed) _The Persian Gulf_ Springer-Verlag, Berlin, 33-56.

ILLING, L.V., 1959. Deposition and Diagenesis of some Upper Palaeozoic carbonate sediments in Western Canada. _World Petrol. Congr., Proc., 5th, New York, 1959, Sect.1_ Paper 2.

————, WELLS, A.J., and TAYLOR, J.C.M., 1965. Penecontemporary dolomite in the Persian Gulf. _Soc. Econ. Paleontologists Spec. Publ._ 13, 89-111.

JAMES, N.P., 1972. Holocene and Pleistocene calcareous crust (caliche) profiles; criteria for subaerial exposure, _J. sedim Petrol._ 42, 817-836.

————, and GINSBURG, R.N., 1980. _The seaward margin of Belize Barrier and atoll reefs. Spec. Publ. 3, International Association of Sedimentologists._ Blackwell, Oxford, 191pp

JANUSSEN, V., 1961. Discontinuity surfaces in limestones. Uppsala University geol. Inst. Bull., 40, 221-241.

JEFFERSON, D.P., 1980. Cyclic sedimentation in the Holkerian (Middle Visean) north of Settle, Yorks., Proc. Yorks. geol. Soc., 42, 483 - 504

JOHNSON, G.A.L., and NUDDS, J.R., 1975. Carboniferous coral geochronometers In ROSENBERG, G.D., and RUNCORN , S.K., (eds) Growth rhythms and the history of the earth's rotation Wiley, London, 27-42.

JONES, B., OLDERSHAW, A.E., and NARBONNE, G.M., 1979. Nature and origin of rubbly limestone in the Upper Silurian Read Bay Formation of Arctic Canada. Sedim. Geol. 24: 227-252.

JONES, T.A., 1921. A contribution to the microscopic study of the Carboniferous Limestone of North Wales, Proc. L'pool geol. Soc., 13, 78-96.

KASIG, W., 1980. Dinantian carbonates in the Aachen region, F.R.G. In BLESS, M.J.M., BOUCKAERT, J., and PAPROTH, E., (eds) Pre-permian around the Brabant massif in Belgium, the Netherlands and Germany. Meded. rijks geol. Dienst., 32, 44-52.

KAYE, C.A., 1959. Shoreline features and Quaternary shoreline changes, Puerto Rico, U.S. Geol. Survey Prof. Paper, 317B, 140pp.

KENDALL, A.C., and BROUGHTON, P.L., 1978. Origin of fabrics in speleothems composed of columnar calcite crystals. J. sedim Petrol. 48, 519-538.

KENDALL, C.G., ST.C., and SKIPWITH, P.A., 1968. Recent algal mats of a Persian Gulf Lagoon. J. sedim Petrol. 38, 1040-1058

KERSHAW, S., 1979. Scourcavities beneath stromatoporoids in level-bottom environments. Palaeontological Association Circular. 98a, 7. (Abstract.)

KIMBERLEY, M.M., 1979. Origin of oolitic iron formations. J. sedim Petrol. 49, 111-132.

KLAPPA, C.F., 1979. Lichen stromatolites: criterion for subaerial exposure and a mechanism for the formation of laminar calcretes (caliche) J. sedim Petrol., 49, 387-400.

KLEIN, G. de V., 1971. A Sedimentary model for determining paleotidal range. Geol. Soc. Am. Bull., 82, 2585-2592.

KNAUTH, P.L., 1979. A model for the origin of chert in limestone. Geology, 7, 274-277.

KOBLUK, D.R., and RISK, M.J., 1977. Calcification of exposed filaments of endolithic algae, micrite envelope formation and sediment production. J. sedim.Petrol.;, 47, 517-528.

KOCURKO, M.J., 1979. Dolomitisation by spray-zone brine seepage, San Andros, Colombia. J. sedim. Petrol., 49,209-214.

KRAUSKOPF, K.B., 1967. Introduction to geochemistry. Mc.Graw-Hill, New York, 721pp.

KRUMBEIN, W.C., and GARRELS, R.H., 1952. Origin and classification of chemical sediments in terms of pH and oxidation-reduction potentials. J.Geol., 60, 1-33.

LAMPLUGH, G.W., 1902. Calcrete. Geol. Mag., 9, 75.

LAND, L.S., 1967. Diagenesis of skeletal carbonates. J.sedim Petrol., 37, 914-930.

LAND, L.S., 1976. Early dissolution of sponge spicules from reef sediments, North Jamaica. J. sedim. Petrol.,46, 967-969

LAPORTE, L.F., 1975. Carbonate tidal-flat deposits of the early Devonian Manlius Formation of New York State. In GINSBURG, R.N. (ed) Tidal Deposits,Springer-Verlag, New York, 243-250.

LINS,T.W., 1976. A discussion of calcisphere ultrastructure and skeletal aragonite from the alga *Acetabularia antillana*. J. sedim. Petrol., 46, 443.

LOEBLICH, A.R., and TAPPAN, H., 1964. Protista, part C, In MOORE, R.C., (ed) Treatise on Invertebrate Paleontology. 900pp.

LOGAN, B.W., 1961. *Cryptozoon* and associated stromatolites from the Recent, Shark Bay, Western Australia. J.Geol., 69, 517-533.

———, 1974. Inventory of diagenesis in Holocene-Recent carbonate sediments, Shark Bay, Western Australia. In LOGAN,B.W., (ed) Evolution and diagenesis of Quaternary carbonate sequences, Shark Bay, Western Australia, Am.Ass. petrol. Geologists Memoir, 22, 195-249

———, HOFFMAN, P., and GEBELEIN, C.D., 1974a. Algal mats, cryptalgal fabrics and structures, Hamelin Pool, Western Australia. In LOGAN, B.W. (ed) Evolution and diagenesis of Quaternary carbonate sequences, Sharkd Bay, Western Australia. Am. Assoc. petrol. Geologists Memoir, 22 140-194.

LOGAN, B.W., READ, J.F., HAGAN, G.M., HOFFMAN, P., BROWN, R.G., WOODS, P.J. amd GEBELEIN. C.D., 1974 b. Evolution and diagenesis of Quaternary carbonate sequences, Shark Bay, Western Australia. Am. Assoc. petrol. Geologists Memoir, 22, 358pp.

———, REZAK, R., and GINSBURG, R.N., 1964. Classification and environmental significance of algal stromatolites. J. Geol., 72, 68-83.

——— , and SEMENIUK, V., 1976. Dynamic metamorphism; processes and products in Devonian carbonate rocks, Canning Basin, Western Australia: Geol.Soc. Australia Spec. Publ., 6, 138pp.

LONGMAN, M.W., 1977. Factors controlling the formation of microspar in the Bromide Formation. J. sedim. Petrol.,47, 347-350

LOWE, D.R., 1975. Water escape structures in coarse-grained sediments. Sedimentology, 22, 157-204

McKINNEY, F.K. and GAULT, H.W., 1980. Palaeoenvironment of late Mississippian fenestrate bryozoans, eastern United States. Lethaia, 13, 127-146.

MAMET, B., and ROUX, A., 1974. Sur quelques algues tubulaires scalariformes de la Tethys Paleozoique. Revue de Micropaleontologie, 17, 134-156.

MARSZALEK, D.S., 1975. Calcisphere ultrastructure and skeletal aragonite from the alga Acetabularia antillana. J. sedim. Petrol., 45, 266-271.

MARSZALEK, D.S., 1976. Calcisphere ultrastructure and skeletal aragonite from the alga Acetabularia antillana: A reply. J. sedim. Petrol. 46, 442.

MATTER, A., 1967. Tidal flat deposits in the Ordovician of western Maryland. J. sedim. Petrol., 37, 601-609.

MATTHEWS, R.K., 1966. Genesis of Recent lime mud in southern British Honduras. J. sedim. Petrol., 36, 428-454.

———, 1971. Diagenetic environments of possible importance to the explanation of cementation fabric in subaerially exposed carbonate sediments. In BRICKER, O.P. (ed) Carbonate Cements, Johns Hopkins, Baltimore, 127-132.

MEYERS, W.J., 1977. Chertification in the Mississippian Lake Valley Formation, Sacramento Mountains, New Mexico. Sedimentology, 24, 75-105.

———, and LOHMAN, K.C., 1978. Microdolomite-rich syntaxial cements: proposed meteoric-marine mixing zone phreatic cements from Mississippian Limestones, New Mexico. J. sedim. Petrol., 48, 475-488

MILLER, J.P., 1952. A portion of the system calcium-carbonate-carbon dioxide-water, with geological implication. Am.J. Sci., 250, 161-203

MILLER, M.H., and WILDING, L.P., 1972. Micromorphological features of soils in relation to plant growth. In PROTZ, R. (ed) Microfabrics of soil and sedimentary deposits. University of Guelph, Ontario, 75-102.

MILLIKEN, K.L., 1979. The Silicified Evaporite Syndrome - Two Aspects of Silicification History of Former Evaporite Nodules from Southern Kentucky and Northern Tennessee, J. sedim. Petrol., 49, 245 - 256

MITCHELL, M., 1978. Dinantian. In MOSELEY, F. (ed) The geology of the Lake District. Yorks. geol. Soc. Occasional Publ.3 Leeds, 168-177.

MOORE C.H.Jr., 1971. Beachrock cements, Grand Cayman Island, B.W.I. In BRICKER, O.P., (ed) Carbonate cements, Johns Hopkins, Baltimore, 9-12.

MOORE, D.G., and SCRUTON, P.C., 1957. Minor internal structures of some Recent unconsolidated sediments. Am. Ass. petrol. Geologists, Bull, 41, 2743-2751.

MORTON, G.H., 1870. The Mountain Limestone of Flintshire and part of Denbighshire, Geol. Mag., 7, 526-527.

———, 1878. The Carboniferous Limestone and Millstone Grit of North Wales: the country between Llanymynech and Minera. Proc. L'pool. geol. Soc. 3, 152-205.

———, 1879. The Carboniferous Limestone and Cefn-y-Fedw Sandstone of the country between Llanymynech and Minera, North Wales, David Bogue, London, 140pp.

———, 1886. The Carboniferous Limestone and Cefn-y- Fedw Sandstone of Flintshire. Proc. L'pool. geol. Soc., 4, 169-208.

NEAVERSON, E., 1929. Faunal horizons in the Carboniferous Limestone of the Vale of Clwyd. Proc. L'pool. geol. Soc., 15, 111-133.

———, 1946. The Carboniferous Limestone Series in North Wales: conditions of deposition and interpretation of its history. Proc. L'pool. geol. Soc., 19, 113-144.

NEUMANN, A.C., GEBELEIN, C.D., and SCOFFIN, T.P., 1970. The composition, structure and erodability of subtidal mats, Abaco, Bahamas. J. sedim. Petrol., 40, 274-297.

———, and LAND, L.S., 1975. Lime mud deposition and calcareous algae in the Bight of Abaco, Bahamas: A budget. J. sedim. Petrol., 45, 763-786.

NEWELL, N.D., IMBRIE, J., PURDY, E.G., and THURBER, D.L., 1959. Organism communities and bottom facies, Great Bahama Bank. Bull. Am. Mus. Nat. Hist., 117, 177-228

———, PURDY, E.G., and IMBRIE, J., 1960. Bahamian oolitic sand, J. Geol., 68, 481 - 497.

NICHOLS, R.A.H., 1966. Petrology of an irregular-nodule bed, Lower Carboniferous, Anglesey, North Wales, Geol. Mag., 103, 477-486.

OLDERSHAW, A.E., 1969. Carboniferous lithofacies, Halkyn Mountain, North Wales, Geol. J., 6, 185-192.

———, and SCOFFIN, T.P., 1967. The source of ferroan and non-ferroan calcite cements in the Halkyn and Wenlock Limestones, Geol. J., 5, 309-320.

PALMER, T.J., and PALMER, C.D., 1977. Faunal distribution and colonization strategy in a Middle Ordovician hardground community. Lethaia, 10, 179-199.

PARK, W.C., and SCHOT, E.H., 1968. Stylolites: Their nature and origin. J. sedim. Petrol., 38, 175-191.

PERKINS, R.D., and ENOS, P., 1971. Hurricane Betsy in the Florida-Bahama area - geologic effects and comparison with Hurricane Donna. J. Geol., 76, 710-717.

PETRYK, A.A., and MAMET, B.L., 1972. Lower Carboniferous algal microflora Southwestern Alberta. Can.J. Earth Sci., 9, 767-802.

PINGITORE, N.E. Jr., 1971. Submarine precipitation of void-filling needles in Pleistocene coral. In BRICKER, O.P.(ed) Carbonate cements, Johns Hopkins, Baltimore, 68-71.

POWER, G., and SOMERVILLE, I.D., 1976. A preliminary report on the occurrence of minor sedimentary cycles in the Middle White Limestone (D_1, Lower Carboniferous) of North Wales. Proc. Yorks. geol. Soc., 40, 491-497

PRATT, B.R., 1979. Early cementation and lithification in intertidal cryptalgal structures, Boca Jewfish, Bonaire, Netherlands Antilles. J. sedim. Petrol., 49, 379-386.

PULLAN, R.A., 1967. A morphological classification of lateritic ironstones and ferruginised rocks in northern Nigeria. Nigerian J. Sci., 1, 161-174.

PURDY, E.G., 1963. Recent calcium carbonate facies of the Great Bahama Bank, 2. Sedimentary facies. J. Geol., 71, 472-497.

PURSER, B.H., 1969. Syn-sedimentary marine lithification of Middle Jurassic limestones in the Paris Basin. Sedimentology, 12, 205-230.

———, 1975. Tidal sediments and their evolution in the Bathonian carbonates of Burgundy, France. In GINSBURG, R.N. (ed). Tidal Deposits, 335-343, Springer-Verlag, New York.

———, and SEIBOLD, E., 1973. The principal environmental factors influencing Holocene sedimentation and diagenesis in the Persian Gulf. In PURSER, B.H. (ed). The Persian Gulf. Springer-Verlag, Berlin, 1 - 10.

de RAAF, J.F.M., BOERSMA, J.R., and van GELDER, A., 1977. Wave-generated structures and sequences from a shallow marine succession, Lower Carboniferous, County Cork, Ireland. Sedimentology, 24, 451-483.

RACZ, L., 1964. Carboniferous calcareous algae and their associations in the San Emiliano and Lois Ciguera Formations (Prov. Leon, N.W. Spain) Leidse. Geol. Meded., 31, 1-112.

RAMSBOTTOM, W.H.C., 1973. Transgression and regression in the Dinantian: a new synthesis of British Dinantian Stratigraphy. Proc. Yorks. geol. Soc., 39, 567-607.

———, 1974. Dinantian. In RAYNER, D.H. and HEMINGWAY, J.E. (eds) The geology and mineral resources of Yorkshire. Yorks. Geol Soc. occasional Publ., 2, Leeds. 45-87.

———, 1977. Major cycles of transgression and regression (mesothems) in the Namurian. Proc. Yorks. geol. Soc., 41, 261-291.

———, 1979. Rates of transgression and regression in the Carboniferous of N.W. Europe. Q. Jl. geol. Soc. Lond., 136, 147-153.

RAUP, D.M., and STANLEY, S.M., 1978. Principles of Palaeontology (2nd ed) Freeman, San Francisco. 1481pp.

READ, J.F., 1973. Carbonate cycles, Pillara Formation (Devonian), Canning Basin, Western Australia, Bull. Can. Petrol. Geol., 21, 38-51.

———, 1974a. Carbonate bank and wave-built platform sedimentation, Edel Province, Shark Bay, Western Australia: Am. Ass. petrol. Geologists. Mem., 22, 1-60.

READ, J. F., 1974b. Calcrete deposits and Quaternary sediments, Edel Province, Shark Bay, Western Australia. Ibid, 250 - 282

———, 1975. Tidal flat facies in carbonate cycles, Pillara Formation (Devonian, Canning Basin, Western Australia. In GINSBURG, R.N., (ed). Tidal Deposits. Springer-Verlag, New York, 251-256.

READ, J.F., and GROVER, G.A., 1977. Scalloped and planar erosion surfaces, middle Ordovician Limestones, Virginia: analogues of Holocene exposed karst or tidal rock platforms. J. sedim. Petrol., 47, 956- 972.

REINECK, H.E., and SINGH, I.B., 1972. Genesis of laminated sand and graded rhythmites in storm-sand layers of shelf mud. Sedimentology 18, 123 - 128.

REVELLE, R., and EMERY, K.O., 1957. Chemical erosion of beach rock and exposed reef rock. U. S. Geol. Survey Profess. Paper 260. T.

RICH, M., 1967. *Donezella* and *Dvinella*, widespread algae in lower and middle Pennsylvanian rocks in east-central Nevada and west-central Utah. J. Paleont., 41, 973-980.

RICHARDSON, P.D., 1968. The generation of scour marks near obstacles. J. sedim. Petrol., 38, 965-970.

RICHTER, D.K. and FUCHTBAUER, H., 1978. Ferroan calcite replacement indicates former magnesian calcite skeletons. Sedimentology, 25, 843-860.

RIDING, R.E., 1970. Palaeoecology and petrology of Middle Carboniferous limestones and stratigraphy of the associated sedimentary sequence, south central Cantabrian Mountains, Spain. Thesis, Univ. Sheffield, 215pp (unpublished)

———, and JANSA, L.F., 1974. *Uraloporella* Korde in the Devonian of Alberta. Can. J. Earth Sci., 11, 1414-1426.

ROBBIN, D.M., and STIPP, J.J., 1979. Depositional rate of laminated soilstone crusts, Florida Keys. J.sedim. Petrol., 49, 175 - 180.

ROGNON, P., 1980. Pluvial and arid phases in the Sahara: the role of non-climatic factos. In ZINDEREN BAKKER, E.M. Van, and COETZEE, J.A., (eds) Palaeoecology of Africa and the surrounding Islands, 12. Balkema, Rotterdam, 45 - 62.

———, and WILLIAMS, M.A.J., 1977. Late Quaternary climatic changes in Australia and the Sahara: a preliminary interpretation. Palaeogeogr. Palaeoclim., Palaeoecol., 21, 285 - 327.

ROSE, W.C.C., and DUNHAM, K.C., 1977. Geology and haematite deposits of south Cumbria. Geol. Surv. U.K. Econ. Mem. London, 170pp..

ROSENBERG, G.D., and RUNCORN, S.K., (eds) Growth rhythms and the history of the earth's rotation. Wiley, London.

RUPP, A., 1967. Origin, structure and environmental significance of Recent and fossil calcispheres. Geol. Soc. Am. spec. Pap., 101, 186 (abstract).

SANDBERG, P.A., 1975. New interpretations of Great Salt Lake ooids and of ancient non-skeletal carbonate mineralogy. Sedimentology, 22, 497 - 537.

SARGENT, H.C., 1912. On the origin of certain clay-bands in the limestone of the Crich inlier, Geol. Mag., 9, 406-412.

SCHENK, P.E., 1967. The Macumber Formation of the Maritime Provinces Canada - a Mississippian analogue to Recent strand-line carbonates of the Persian Gulf. J. sedim. Petrol., 37, 365 - 376.

———, 1975. Carbonate-sulfate intertidalites of the Windsor Group (Middle Carboniferous), Maritime provinces, Canada. In GINSBURG, R.N. (ed) Tidal Deposits, Springer Verlag, New York, 373 - 380.

SCHNEIDER, J.F., 1975. Recent tidal deposits, Abu Dhabi, U.A.E., Arabian Gulf In GINSBURG, R.N. (ed) Tidal Deposits, Springer -Verlag, New York, 209 - 214.

SCHNEIDERMANN, N., SANDBERG, P.A., and WUNDER, S.J., 1977. Recognition of early cementation of aragonitic skeletal carbonates. Nature phys. Sci., Lond. 240, 88-89.

SCHOPF, T.J.M., 1980. Paleoceqnography. Harvard University, Harvard University, Cambridge, Massachusetts, 341 pp

SCHWARZACHER, W., 1958. The stratification of the Great Scar Limestone in the Settle district of Yorkshire. L'pool. Manch'r. geol. J., 2, 124- 142.

———, 1961. Petrology and structure of some Lower Carboniferous reefs in northwestern Ireland. Am. Ass. petrol. Geologists. Bull., 45, 1481- 1503.

———, 1975. Sedimentation models and quantitative stratigraphy. Developments in Sedimentology, 19, Elsevier, Amsterdam, 382pp.

SCOFFIN, T.P., 1970. The trapping and binding of subtidal carbonate sediments by marine vegetation in Bimini Lagoon, Bahamas. J. sedim. Petrol., 40, 249-273.

SEIBOLD, E. 1980. Climate indicators in marine sediments off north west Africa - A critical review In ZINDEREN BAKKER, E.M. Van, and COETZEE, J.A., (eds) Palaeoecology of Africa and the surrounding Islands, 12, Balkema, Rotterdam. 175 - 187

SELLEY, R.C., 1970. Studies of sequence in sediments using a simple mathematical device Q. Jl. geol. Soc. Lond., 125, 557-581.

SHEARMAN, D.J., TWYMAN, J., and ZAND KARIMI, M., 1970. The genesis diagenesis of oolites. Proc. Geologists Assoc., 81, 561 - 575.

SHINN, E.A., 1968. Practical significance of birdseye structures in carbonate rocks. J. sedim. Petrol., 38, 215 - 223.

———, GINSBURG, R.N., and LLOYD, R.M., 1965. Recent supratidal dolomite from Andros Island, Bahamas. In PRAY, L.C. and MURRAY, R.C. (eds). Dolomitisation and Limestone Diagenesis - a symposium. Soc. Econ. Paleont. Mineral. Spec. Publ., 13, 112 - 123.

———, LLOYD, R.M., and GINSBURG, R.N., 1969. Anatomy of a modern carbonate tidal-flat, Andros Island, Bahamas. J. sedim. Petrol., 39, 1202 - 1228.

SIMS, J.D., 1975. Determining earthquake recurrence intervals from deformational structures in young lacustrine sediments. Tectonophysics, 29, 141 - 152.

SLEEMAN, J.R., 1963. Cracks, peds and their surfaces in some soils of the Riverine Plain, N.S.W., Aust. J. Soil Res., 1, 91 - 102

SMOSNA, R., and WARSHAUER, S.M., 1978. A scheme for multivariate analysis in carbonate petrology, with an example from the Silurian Tonoloway Limestone. J. sedim Petrol., 49, 257 - 272.

SMITH, A.G., BRIDEN, J.C., DREWRY, G.E., 1973. Phanerozoic World Maps. In HUGHES, N.F., (ed). Organisms and continents through time. Spec. Pap. Palaeontology, 12, 1 - 42.

SMITH, S., 1915. The genus Lonsdaleia and Dibunophyllum rugosum (Mc.Coy) Q. Jl. geol. Soc. Lond., 71, 218 - 272.

SOMERVILLE, I.D., 1977. The sedimentology and stratigraphy of the Dinantian limestones in the Llangollen area and east of the Clwydian range, North Wales. Unpublished Ph.D. dissertation. Queen's University of Belfast, 234 pp.

——— , 1979a. Minor sedimentary cyclicity in late Asbian (Upper D_1) limestones in the Llangollen district of North Wales. Proc. Yorks. geol. Soc., 42, 317 - 341.

——— , 1979b. A sedimentary cyclicity in early Asbian (Lower D_1) limestones in the Llangollen district of North Wales, Proc. Yorks. geol. Soc., 42, 397 - 404

——— , 1979c. A cyclicity in the early Brigantian (D_2) limestones of the Clwydian Range, North Wales, and its uses in correlation. Geol. J., 14, 69 - 86.

SORBY, H.C., 1856. On the origin of the Cleveland Hill ironstone. Proc. geol. polytech. Soc. Wes. Riding Yorks., 3, 457- 461.

STACH, L.W., 1936. Correlation of zooarial form with habitat. J. Geol., 44, 60 - 65.

STANLEY, D.J., and MALDONADO, A., 1979. Levantine Sea - Nile Cone lithostratigraphic evolution: quantitative analysis and correlation with palaeoclimatic and eustatic oscillations in the late Quaternary. Sedim. Geol., 23, 37 - 65.

STANTON, R.O., 1963. Upper Devonian calcispheres from Redwater and South Sturgeon Lake reefs, Alberta, Canada. Bull. Can. petrol. Geologists, 11, 410 - 418.

STEVENSON, I.P., and GAUNT, G. D., 1971. Geology of the country around Chapel en le Frith, Geol. Survey, Great Britain (Mem), 444 pp.

STOCKMAN, K.W., GINSBURG, R.N., and SHINN, E.A., 1967. The production of lime-mud by algae in south Florida, J. sedim. Petrol., 37, 633 - 648.

STODDART, D.R., and CANN, J. R., 1969. Nature and origin of beach rock. J. sedim. Petrol., 35, 243 - 247.

STRATTON, J.F., and HOROWITZ, A.S., 1974. Studies of the flow of water through models of Polypora. Docum. Lab. Geol. Fac. Sci. Lyon. H.S. 3 (fasc. 2), 425 - 438.

SWEETING, M.M., 1966. The weathering of limestones. In DURY, G.H. (ed) Essays in geomorphology. Heinemann, London, 171-210.

TAYLOR, J.C.M., and ILLING, L.V., 1971. Variation in Recent beachrock cements, Quatar, Persian Gulf. In BRICKER, O.P. (ed). Carbonate cements, Johns Hopkins, Baltimore, 32 - 33.

TEBBUT, G.E., CONLEY, C.D., and BOYD, D.W., 1965. Lithogenesis of a distinctive carbonate rock fabric. In PARKER, R.B. (ed) Contributions to geology. University Wyoming, Laramie, 1 - 13.

THOMAS, T.M., 1953. New evidence of intra-formational piping at two separate horizons in the Carboniferous Limestone (Dibunophyllum zone) at South Cornelly, Glamorgan. Geol. Mag. 90, 73 - 82.

THOREZ, J. 1976, Practical identification of clay minerals. Lelotte, Dison, 90pp.

TIETZ, G., and MULLER, G., 1971. High-magnesian calcite and aragonite cementation in Recent beachrocks, Fuerteventura, Canary Islands, Spain. In BRICKER, O.P. (ed) Carbonate cements. Johns Hopkins, Baltimore, 4 - 5.

TOOMEY, D.F., and JOHNSON, J.H., 1968. Ungdarella americana a new red alga from the Pennsylvanian of southeastern New Mexico. J. Paleont., 42, 556- 560.

TRICART, J., 1972. Landforms of the humid tropics, forests and savannas (Translation from French by de Jonge, C.J.K) Longman, London, 306pp.

TUCKER, M. E., 1973. Sedimentology and diagenesis of Devonian pelagic limestones (Cephalopodenkalk) and associated sediments of the Rhenohercynian Geosyncline, West Germany, Neues Jb. Geol. Palaont. Abh., 142, 320 - 350.

———, 1974. Sedimentology of Palaeozoic pelagic limestones: the Devonian Griotte (southern France) and Cephalopodenkalk (Germany). In HSU, K.J. and JENKYNS, H.C. (eds) Pelagic sediments on land and under the sea. Spec. publ. int. Ass. Sediment., 1, 71 - 92

———, 1975. Vadose diagenetic fabrics in Triassic lacustrine limestones from South Wales. Extraits des publications du congres 1Xme Congres International de Sedimentologie, Nice, 217 - 223.

VAUGHAN, A., 1905. The palaeontological sequence in the Carboniferous Limestone of the Bristol area. Q. Jl. geol. Soc. Lond., 61, 181 - 307.

WAGNER, C.W., and TOGT, C. van der, 1973. Holocene sediment types and their distribution in the southern Persian Gulf. In PURSER, B.H. (ed) The Persian Gulf. Springer Verlag, Berlin, 123 - 156.

WALKDEN, G.M., 1970. Environmental studies in the Carboniferous Limestone of the Derbyshire Dome. Unpubl. Phd. dissertation, University of Manchester, 2 vols., 182pp.

WALKDEN, G.M., 1972. The mineralogy and origin of interbedded clay wayboards in the Lower Carboniferous of the Derbyshire Dome, Geol. J., 8, 143 - 159.

———, 1974, Palaeokarstic surfaces in Upper Visean (Carboniferous) Limestones of the Derbyshire Block, England, J. sedim. Petrol., 44, 1232 - 1247.

———, 1977. Volcanic and erosive events on an Upper Visean carbonate platform, North Derbyshire. Proc. Yorks. geol. Soc., 41, 347.

WALKER, R.G., 1979. Facies Models, 211pp. Geoscience, Canada, reprint series 1, Toronto.

WALKER, K.R., 1972. Community ecology of the Middle Ordovician Black River Group of New York State. Geol. Soc. Am. Bull. 83, 2499 - 2524.

WALLS, R.A., HARRIS, W.B., and NUNAN, W.E., 1974. Calcareous crust (caliche) profiles and early subaerial exposure of Carboniferous carbonates, north-eastern Kentucky. Sedimentology, 15, 417 - 440.

WALTHAM, A.C., 1971. Shale units in the Great Scar Limestone of the southern Askrigg Block. Proc. Yorks. geol. Soc., 38, 285 - 292.

WANLESS, H.R., 1979. Limestone response to stress: pressure solution and dolomitisation. J.sedim. Petrol., 49, 437 - 462.

———, and WELLER, J.M., 1932. Correlation and extent of Pennsylvanian cyclothems. Geol. Soc. Am. Bull., 43, 1003 - 1016.

WARDLAW, N., OLDERSHAW, A. and STOUT, M. 1978. Transformation of aragonite to calcite in a marine gastropod. Can. J. Earth Sci., 15, 1861-1866.

WEDD, C.B., SMITH, B., and WILLS, L.J., 1927. The geology of the country around Wrexham, Part 1. Geol. Surv. Great Britain Memoir, 121, 179pp

———, ———, KING, W.B.R., and WRAY, D.A., 1929. The country around Oswestry. Geol. Surv. Great Britain Memoir 137, 234pp.

WELANDER, P., 1961. Numerical prediction of storm surges. Advances in Geophys., 8, 315 - 379. Academic Press, New York.

WELLS, A.J., 1956. ;The development of chert between the Crow and Main Limestones in North Yorkshire. Proc. Yorks. geol. Soc., 30, 177 - 196.

WIGLEY, T.M.L., and PLUMMER, L.N., 1976. Mixing of carbonate waters. Geochim.Cosmochim.Acta, 40, 989 - 995.

WILSON, J.L., 1975. Carbonate facies in geologic history. Springer-Verlag, New York. 471pp.

WILSON, J.L. ., 1967. Carbonate evaporite cycles in the lower Devonian Duperow Formation of Williston Basin. Bull. Can.petrol. Geol. , 15, 230 - 312.

WOOD, A., 1940. Two new calcareous algae of the family Dasycladaceae from the Carboniferous Limestone Limestone Proc. Lpool. geol. Soc., 18, 14-18.

WOODS, P.J., and BROWN, R.G., 1975. Carbonate sedimentation in an arid zone tidal flat, Nilemah Embayment, Shark Bay, Western Australia. In GINSBURG, R.N., (ed) Tidal Deposits. Springer-Verlag, New York.

WRAY, J.L., 1977. Calcareous Algae. Developments in Palaeontology and Stratigraphy No. 4. Elsevier, Amsterdam, 185pp.

ZENGER, D.H., 1972. Dolomitisation and Uniformitarianism. J. Geol. Education, 20, 107 - 124.

APPENDICES

APPENDIX 1.

SPONGE AFFINITIES OF SOME CARBONIFEROUS CHAETETIDS.

During the investigation of the Tynant Formation (Lower Asbian) fauna a small number of chaetetid colonies were discovered.

Acetate peels of these colonies revealed clear calcitic streaks within their calcite walls. The resemblance of these streaks to spicules was investigated through further sectioning and S.E.M microscopy. Previous literature had reported calcitic spicules within Mesozoic chaetetids (Dieci et al, 1977), but the variability and poor preservation of these Lower Carboniferous microstructures remained problematic to their spicular interpretation.

Comparison with material in the British Museum (Natural History), Merseyside County Museum, and Royal Scottish Museum led to the discovery of further spicule-bearing specimens, of comparable age from North Wales, northern England and southern Scotland. Among these were two silicified specimens (collected from Eglwyseg, and probably from chert nodules within the Tynant Formation) that contained exceptionally well preserved microgranular silica spicule pseudomorphs (of monaxon tylostyle form), embodied within chalcedonic silica walls.

From comparison with extant sclerosponge material and petrographic study of the calcareous skeleton, it was concluded that the original microstructure of this spicular chaetetid, identified as Chaetetes (Boswellia) mortoni sp. nov. was probably aragonite.

The inter-relationships of spicule preservation, and detail retained by the siliceous spicule pseudomorphs suggested a primary mineralogy of opal 'A' for the spicules.

The spicular nature of this Upper Palaeozoic chaetetid confirms views of Hartman and Goreau (1972) that at least some (Palaeozoic) chaetetids are sponges.

A copy of the published report on these chaetetids is submitted with this thesis.

APPENDIX II

POINT COUNT ANALYSES OF ASBIAN CARBONATES

The following tables list the point count percentage analyses for Asbian carbonates in the Llangollen and Oswestry areas. Point count analyses of Potts Beck Limestone samples are also given.

Table	A	Tynant Formation
	B	Tynant Formation
	C	Tynant Formation
	D	Sychtyn Member (Whitehaven Formation)
	E	Whitehaven Formation
	F	Llanymynech Member (Whitehaven Form'n)
	G	Eglwyseg Formation.
	H	Eglwyseg Formation
	I	Potts Beck Limestone (Ravenstonedale Trough)

KEY TO ROW NUMBERS GIVEN ON TABLE A.

TYNANT FORMATION

TABLE A

Specimen / Microfacies		16 / 2.2.	25 / 2.1.	6 / 2.1.	17 / 2.1.	4 / 2.1.	29 / 2.1.	191 / 2.1.	1036 / 3.1	1038 / 3.1
		%	%	%	%	%	%	%	%	%
Lime mud+ bioclastic silt	1	64	46.7	65.0	38.5	21.5	33.8	24.0	0	0
Peloids + intraclasts +?.	2	3.3	5.2	1.8	3.3	5.0	6.0	15.8	22.5	30.5
Burrow disturbance + internal sediment	3	0	0	0	0	0	0	0	0	0
Cement + fenestra druse	4	0	0.8	0.3	0.3	0	0	2.3	19.8	27.8
Brachiopod	5	3.5	1.2	0.7	2.3	2.8	1.3	0.5	0.2	0
Coral	6	0	0	0	0	0	0	0	0	0
Mollusc	7	0	0.7	1.3	1.5	0.5	0	0.8	0	0
Bryozoa	8	0	0	0	0	0	0	0	0	0
Echinoderm	9	1.0	0.3	1.3	0	0	0.8	8.0	0	0
Thick,irregular-walled calcispheres	10	0.5	0.3	1.8	0	0	0	0	1.0	2.2
Smooth thin walled calcispheres	11	2.5	3.5	1.2	3.3	2.0	2.3	1.5	4.5	3.2
Beresellids + other tubular septate ?algae	12	19.2	35.4	19.2	44.3	64.8	51.3	40.0	46.0	26.2
Koninkopora	13	0	0	0.8	0	0.5	0	1.0	0	0
Ungdarellid + other red algae + other algae.	14	2.0	2.0	2.2	0.8	0.5	0	0.8	0.2	0
Foraminifera	15	0.7	3.0	2.8	4.0	1.0	3.5	5.3	4.5	9.8
Ostracod	16	2.8	1.3	1.3	2.0	1.5	1.3	3.0	1.3	0.5
$\frac{7+10+11}{5+6+\dots+15}$ C	17	10.2	9.8	13.8	8.5	3.2	3.8	3.1	9.8	13.1
$\frac{12+13+14}{5+6+\dots+15}$ A	18	72.2	80.5	70.8	80.4	91.9	86.9	72.6	81.9	63.3
$\frac{5+6+8+9+15}{5+6+\dots+15}$ M	19	17.6	9.7	15.4	11.2	4.9	9.3	24.3	8.3	23.5
Number of counts		600	600	600	400	400	400	400	600	600

- TYNANT FORMATION

TABLE B

Specimen	1043	1046	1047	1048	1480	1485	1486	1484	1482	1483
Microfacies	1.1	1.1	1.1	1.1	1.1	1.1	1.1	2.1	3.1	3.1
Character (see table A)	%	%	%	%	%	%	%	%	%	%
1	42.3	54.3	60.3	59.0	67.3	66.8	79.5	51.0	12.7	10.2
2	19.9	13.3	12.0	7.3	2.2	3.7	2.3	7.2	17.3	12.8
3	0	0	5.0	3.2	2.3	2.5	1.7	0	0	0
4	0	0	0	0	0	0	0	0.5	14.2	9.7
5	0	0	0	0	0.2	0	0	0	0.5	0.2
6	0	0	0	0	0	0	0	0	0	0
7	0	0	0	0	2.2	,0	0	0.2	0.2	0
8	0	0	0	0	0	0	0	0	0	0
9	0	0	0	0	0	0	0	0	0	0.2
10	9.8	13.9	15.1	25.2	9.7	12.8	7.8	9.8	1.0	0.7
11	2.8	8.8	1.1	1.2	0.8	2.3	1.2	2.3	4.0	3.3
12	15.0	7.5	4.9	2.7	10.3	9.3	6.8	26.2	40.0	57.0
13	0	.0	0	0	0	0.2	0	0	0.3	0.8
14	1.7	0	0	0.3	1.7	0	0	0	0	0.3
15	0,	1.5	0.3	0.5	1.3	0.2	0	1.3	7.8	3.0
16	1.3	0.7	1.8	1.2	1.2	2.0	0.7	1.2	1.5	1.7
17	42.9	75.9	75.7	88.3	48.4	61.0	56.9	31.0	9.6	6.1
18	57.1	20.0	22,7	10.0	45.9	38.3	43.2	65.7	74.9	88.8
19	0	4.0	1.5	1.7	5.7	0.7	0	3.3	15.5	5.1
Number of counts	All 600 counts.									

TYNANT FORMATION TABLE C

Specimen	1037	1039	1040	(1042)	1044	1045	1478	1479	1481
Microfacies	3.1	2.1	2.1	2.1/1.1	3.1	3.1	2.1	2.1	2.1
Form^n/ Member	All Tynant Formation								
Character (see table A)	%	%	%	%	%	%	%	%	%
1	0	41.9	29.8	24.4	0	0.5	43.5	30.8	34.5
2	38.4	21.1	15.0	27.1	20.7	11.4	5.7	17.0	5.7
3	0	0	0	0	0	0	0	0	0
4	29.6	0	3.8	0	19.2	22.7	0	0	0
5	0	0	9.0	2.0	0.2	0	2.3	0.3	1.2
6	0	0	4.8	0	0	0	1.0	0.2	0
7	0	0	4.3	1.0	1.2	0.5	0	0	0.2
8	0	0	0	0	0	0	0	0	0
9	0	0	0.2	0	2.0	0	0	0	0
10	3.3	0.6	1.3	3.0	0.2	0.2	1.7	2.3	2.5
11	4.2	3.7	3.8	5.5	5.5	2.2	4.3	4.2	4.2
12	16.0	29.5	17.2	32.6	41.8	53.7	34.8	40.5	40.8
13,	0	0	0	0	0	0	0	0	1.0
14	0	0	5.7	3.7	0.3	0.3	2.2	0	4.3
15	7.0	2.8	2.8	0.4	8.7	8.2	2.2	1.0	4.5
16	1.2	0.4	2.0	0.4	0.2	0.7	0.5	1.3	0.5
17	24.6	11.8	19.3	19.6	11.4	4.4	12.4	13.4	11.7
18	52.4	80.6	46.5	75.5	70.5	83.0	76.3	83.5	78.7
19	22.9	7.5	34.2	4.9	18.1	12.6	11.3	3.1	9.7
Number of counts	600	600	600	600	600	600	600	600	600

SYCHTYN MEMBER

TABLE D.

Specimen	1074	1216	1077	1071	1069	1374	1106	1220	1377	1322
Microfacies	3.3	3.3	1.1	2.1	1.1	3.1	3.1/3.3	3.1	3.1	2.1
Character (see table A for key.	%	%	%	%	%	%	%	%	%	%
1	0	0	88.8	27.3	60.3	0	1.3	1.0	0	8.5
2	63.5	51.0	0.3	11.0	0.3	5.5	43.8	17.8	25.0	20.0
3	0	0	0.3	0	0	0	0	0	0	0
4	18.0	18.0	0.3	1.0	2.3	10.5	17.8	11.8	12.8	4.8
5	0	0	0	2.3	0	1.0	1.3	0	0.3	0.3
6	1.0	0	0	0	0	0	1.8	1.3	0.3	0
7	0	1.0	0	1.0	5.5	0	1.5	0	0	0.
8	0	0	0	0	0	0	0	0	0	0
9	1.8	3.5	0	0.8	0.3	0	0.8	0.5	0.3	1.0
10	0	1.0	5.0	1.5	6.3	0	0	0.5	0.3	0
11	0.8	1.0	2.0	2.5	9.0	1.0	0.3	0	2.0	3.3
12	9.8	17.8	2.0	46.3	14.5	73.8	28.0	57.0	47.0	50.8
13	0.8	0.5	0	1.0	0.8	3.3	1.8	3.0	2.5	0
14	0.8	0.8	0	0	0	1.0	0	0	1.3	0
15	3.5	5.5	0.5	4.3	0.5	0.8	2.0	6.0	7.8	8.8
16	0.3	0	1.0	0.3	0.5	3.3	0	1.3	0.8	2.8
17	4.1	9.7	73.7	8.4	56.5	1.2	4.7	0.7	3.6	4.5
18	61.6	61.3	21.0	79.4	41.6	96.6	80.0	87.9	81.5	81.6
19	34.3	29.0	5.3	12.2	2.0	2.2	15.4	11.4	14.9	13.9
Number of counts	All 400 counts.									

WHITEHAVEN FORMATION

TABLE E

Specimen	1318	1320	1097	1375	1321	1412	1409	1415	1404	1143
Microfacies	3.1	3.1	3.1	1.1	1.1	3.1	3.1	3.3	3.1	3.1
Member	Sychtyn Member					Llanymynech Member				
Character (see table A for key)	%	%	%	%	%	%	%	%	%	%
1	0	0	0	71.3	83.0	0	0	0	0	0
2	30.0	44.0	19.5	1.5	0	24.5	23.0	50.3	16.8	27.0
3	0	0	0	0	0	0	0	0	0	0
4	25.8	21.8	8.8	2.3	0	23.8	27.0	22.3	33.0	23.5
5	0	0.3	0	0.3	0	0.8	2.3	0	0.3	1.5
6	0	0	0	0	0	9.0	0	0	0	0
7	0	1.3	0	0	0	0.8	2.0	0	2.0	0.5
8	0	0	0	0	0	0.3	0.5	0	0	0
9	1.5	1.8	1.5	0	0.3	0	5.8	0.5	0	1.5
10	2.0	0.5	3.0	2.8	6.3	0	0	0.3	0.3	0
11	0.3	0.8	2.0	6.5	5.5	1.0	0.8	0.5	1.0	1.5
12	26.8	22.5	58.0	11.5	3.0	25.8	15.0	20.0	28.3	35.3
13	1.8	0.3	0	0.5	0	4.8	13.5	0.5	8.8	2.0
14	6.3	0	0.8	0	0	0	7.0	1.0	0.8	0
15	5.3	6.8	6.3	0.5	0	7.3	3.3	4.3	7.3	5.8
16	0.5	0.3	0.3	3.0	1.8	1.3	0	0.5	1.8	1.5
17	5.1	7.4	7.0	42.0	78.3	3.5	5.5	2.8	6.7	4.2
18	79.4	66.9	82.2	54.5	20.0	61.6	71.0	79.6	77.8	77.6
19	15.4	25.7	10.8	3.4	1.7	34.8	23.5	17.6	15.5	18.2
Number of counts	All 400 counts									

compositional variation of Llanymynech Member sediments. TABLE F

Specimen	1142	1151	1136	1144	1146	1163	1149	1148	1168	1170	1184	1167
Microfacies	3.1	3.1	3.1	3.1	3.2	3.2	3.4	3.2	3.2	3.2	3.2	3.2
Character (see table for key)	%	%	%	%	%	%	%	%	%	%	%	%
1	0	0	0	0	1.0	0	0	0	0	0	0	0
2	26.5	27.5	24.8	26.3	26.0	22.0	41.0	15.5	12.8	14.8	14.3	18.8
3	0	0	0	0	0	0	0	0	0	0	0	0
4	37.0	39.5	35.3	35.0	12.5	7.3	26.8	39.5	40.0	36.8	43.8	39.0
5	0.8	3.5	1.3	2.3	7.0	2,8	3.3	3.8	1.3	4.3	5.0	7.0
6	0	0	0	0	4.5	0	0.8	0.3	0.8	1,8	1.8	0.3
7	0.5	0	0.5	1.5	0.5	0	1.5	19.0	5.8	7.8	5.0	3.0
8	0.5	2.0	0.8	0	3.8	0	4.5	1.5	14.0	16.3	10.5	18.8
9	0.3	0.5	3.8	6.3	29.8	48.0	15.3	16.5	22.8	14.3	13.8	9.0
10	0.5	0.8	0	0.3	0	0	0	0	0	0	0	0.3
11	2.0	2.0	1.0	1.0	0.5	0.3	0	0	0.3	0.3	0	0.8
12	22.0	14.8	11.8	8.5	4.5	13.5	2.8	0.8	0.5	1.0	1.0	1.3
13	1.0	0.5	14.0	14.0	1.5	0	1.3	1.3	0	1.3	2.5	0
14	0	0	0.3	0.8	0.3	0.8	2.5	0	0	0	1.5	0
15	4.5	5.3	5.0	3.5	4.5	4.8	2.0	2.0	0.3	1.8	0.5	1.5
16	1.5	3.8	1.5	0.8	1.5	0.5	1.0	0.3	0	1.0	0	0.8
17	9.4	9.4	3.9	7.2	1.7	0.4	4.4	42.5	13.3	16.8	12.0	9.6
18	71.9	52.1	68.0	61.2	12.9	20.3	19.3	4.5	1.1	4.7	12.0	3.0
19	18.8	38.5	28.1	31.6	85.3	79.3	76.3	53.1	85.6	78.4	75.9	87.3
Number of counts	All 400 counts.											

EGLWYSEG FORMATION

TABLE G.

Specimen	040	038	044	031i	031iv	033	043	049	039a	031ii	036
Microfacies	3.1	3.1	2.1	2.1	1.1	3.1	2.1	3.1	3.1	1.4/3.1	2.1
Character (see key on table A).	%	%	%	%	%	%	%	%	%	%	%
1	0	0	49.5	61.3	66.8	0	58.5	0	0	46.3	35.7
2	7.3	28.5	1.3	0.5	0.5	10.0	0.5	28.8	36.3	15.0	8.3
3	0	0	0	0	0	0	0	0	0	0	0
4	40.0	26.8	0	0.5	0	46.8	0	23.3	27.8	3.3	4.5
5	1.5	0.5	0.3	0	0	0.5	1.8	0.5	3.3	0	2.3
6	1.3	12.3	0	0	0	0	0	0	5.8	0	2.0
7	9.0	5.0	0.3	0.8	1.8	2.5	0	0	0	0	0
8	0	0	0	0	0	0	0	0	0	0	0
9	4.8	0	3.5	6.8	0	3.5	2.5	0	0.8	0	3.8
10	0	0.5	0	0.3	6.8	0	0	0.8	0.3	0.5	1.3
11	3.0	3.0	5.5	3.5	8.0	3.0	5.3	4.3	1.5	7.5	4.0
12	25.5	18.3	33.5	15.8	9.8	27.8	25.5	39.5	20.0	23.5	27.8
13	0	1.0	0.3	2.5	0.8	0	0	0.8	1.3	0.5	0
14	0	0	0	4.3	0	0	0.5	0	0	0	0.5
15	6.0	4.0	4.8	4.0	4.0	5.5	4.5	2.0	3.3	2.3	9.0
16	1.8	0.3	1.3	0	1.8	0.5	1.0	0.3	0	0.8	0.8
17	23.5	19.1	12.0	12.0	53.2	12.9	13.7	10.5	4.9	23.4	10.4
18	50.0	43.3	70.3	59.3	33.9	64.9	68.0	84.3	59.0	70.1	55.9
19	26.5	37.6	17.7	28.7	12.9	22.2	18.3	5.2	36.1	6.6	33.7
Nº of counts/ specimen	All 400 counts.										

EGLWYSEG FORMATION

TABLE H

Specimen	717	687i	687ii	1596	1570	1246	1254	1241	1234	1538	034	461
Microfacies	3.2	4.1	3.1	4.	1.2a	3.3	3.2	3.1	1.1	3.1	3.2	3.2
Character (see key on table A)	%	%	%	%	%	%	%	%	%	%	%	%
1	0	9.5	37.5	3.0	33.5	0	5.0	3.3	70.8	3.0	0	0
2	20.5	9.5	1.5	15.3	50.5*	52.8	15.5	16.8	5.0*	16.8	14.5	10.5
3	0	0	0	0	0	0	0	0	0	0	0	0
4	34.0	41.5	14.5	45.3	0	35.5	39.0	41.3	3.3	23.0	27.5	42.8
5	0.8	7.0	2.0	32.5	0	0.8	17.5	2.3	0.3	1.3	10.3	1.3
6	0.3	0	1.5	0	0	0	0	0	0	0	4.8	1.3
7	0.3	5.5	6.5	0	0	1.0	1.0	3.3	0	2.0	0.8	1.3
8	1.0	0	0	0	0	0	0	0	0.5	0	0	0.8
9	20.0	20.0	3.5	1.5	0	0.8	9.3	1.8	0.5	9.8	7.8	16.8
10	0	0	0	0	1.3	0.3	0.8	0.3	1.3	0.3	0	0
11	1.5	1.5	3.0	0	4.3	2.8	1.5	2.3	2.0	2.5	2.0	0
12	5.5	5.5	20.0	0	8.5	3.5	5.3	12.3	10.5	4.5	17.5	0
13	8.8	0	1.0	2.0	0	0.3	0.8	2.3	1.5	0.5	2.8	17.0
14	2.5	0	0	0	0	0	0	0	0	1.8	0	3.5
15	4.8	0	8.0	0.3	0.8	2.5	4.0	13.0	3.0	3.8	11.8	4.5
16	0.3	0	1.0	0	1.0	0	0.5	1.3	1.5	0	0.5	0.5
17	3.9	16.7	20.9	0	37.3	34.0	8.1	15.0	16.9	17.4	4.8	2.7
18	37.0	14.1	46.2	5.5	57.6	34.0	15.0	38.9	55.8	28.4	35.2	44.3
19	59.1	69.2	33.0	94.5	5.1	31.9	76.9	45.6	27.3	54.1	60.0	53.0

No of counts/specimen   All 400 counts.

* Lime mud composed of ?faecal pellets.

POTTS BECK LIMESTONE

TABLE I

Specimen	1857a	b	c	d	e	f	g	h	i	j
Microfacies	2.1	3.1	3.1	3.1	3.1	3.1	3.3	1.1	1.1	1.1
Character (see key on table A).	%	%	%	%	%	%	%	%	%	%
1	35.8	7.3	12.3	8.0	3.0	5.0	7.2	74.3	68.0	85.3
2	11.0	29.5	37.3	46.2	46.2	47.5	51.2	1.7	2.0	0.7
3	0	0	0	0	0	0	0	0.3	6.5	6.2
4	0	14.7	19.8	18.0	21.0	16.5	17.5	0	0	0
5	1.3	4.8	3.8	2.5	5.7	2.0	1.5	0.2	0	0.1
6	0	0	1.7	0	0.2	0.7	0	0	0	0
7	0	0	5.2	0.2	0.3	1.0	0.7	0	0.2	0.3
8	0	0	0	0	0	0	0	0	0	0
9	2.8	2.0	1.2	0.8	0.2	1.0	0.8	0	0	0
10	0.5	0.5	0	0.2	0	0	0	2.0	4.8	2.0
11	1.3	2.3	1.2	1.2	0.7	1.0	1.5	2.7	13.2	1.0
12	42.0	25.5	12.8	15.2	14.2	16.2	13.3	17.5	4.2	2.7
13	0	0	0	0	0	0	0	0	0	0
14	3.5	0	0.3	0.2	0.2	0.8	0.5	0	0	0
15	1.3	12.5	4.0	7.2	7.7	8.3	5.7	0.5	0	0.3
16	0.3	0.8	0.3	0.5	0.7	0	0.2	0.8	1.2	1.3
17	3.5	5.9	21.0	5.5	3.5	6.5	9.0	20.5	81.3	51.6
18	86.1	53.5	43.6	56.1	49.4	54.8	57.6	76.6	18.7	41.3
19	10.4	40.6	35.4	38.4	47.1	38.7	33.3	2.9	0	7.1
N° of counts /specimen.	All 600 counts.									

APPENDIX III

X-Ray Diffractograms and Clay Mineral Analyses of Selected Argillites and Carbonates

X-Ray diffractogram analyses of the clay grade (< 2μm) insoluble component of some argillites and carbonates was undertaken to attempt to distinguish the mineral assemblages of subaerially emergent episodes of sedimentation and sediment modification from marine deposits. The assemblage of clay minerals associated with laminated calcrete crusts and tubular calcrete structures is of particular interest due to their association with underlying (marine) carbonates, and overlying subaerial deposits or palaeokarsts.

Technique

Carbonate samples were fragmented, and lightly ground with mortar and pestel (~100g). The resultant sand and powder was then dissolved in 5% acetic acid solution, leaving a minor amount of insoluble residue. Careful decanting allowed replenishment of spent acid.

Weak acid is prefered, to lessen possible reactions with and removal of chlorite minerals.Washing of the resulting residue comprised two or three stage dilutions by decanting upper layers following a settling period of 5 or 6 hours. The residue was then resuspended, and 200ml placed in a measuring cylinder, from which the upper c.50ml was decanted,after a 2minute settling period,into a centrifuge tube, at the base of which a glass X.R.D. mounting plate had been placed. The samples were then centrifuged, and the plates removed and dried.

Specimens were run 'normal', after glycolation, and after heating to 500°C for 4hrs.

Co K_α radiation was used, and specimens run from 3° to 34° 2θto encompass the 001 reflections of clay minerals and the principle quartz reflection.

Whilst it is recognised that quantitative assesment of X.R.D. data of clay minerals is subject to many problems (Thorez,1976) a simple comparative 'semi-quantitative' technique was employed.

Peak areas were measured for the 001 reflections of the 7Å kaolinite peak; 10Å illite peak; 10-14 Å mixed layer illite smectites, and 14Å smectites. Glycolation provided a check on estimation of peak areas where they overlapped. The resulting analyses, however, are, at best, accurate to only one significant figure.

As the peak areas were compared between samples,weighting for individual mineral phases was not undertaken.

The following Table J lists the data, and representative diffractograms are appended (as figs A i to Aiii).

Interbedded shales, not asociated with palaeokarsts, are dominantly illitic, although samples from the thick shale of Tf.7.8, Trefor Formation are dominated by a mixed layer illite-smectite component.(s/m 1395 & 530).

Insoluble residues from carbonates , including those immediately beneath palaeokarstic surfaces (e.g. s/m1102)(beneath Asbian Brigantian boundary at Nant Mawr) are also dominantly illitic, although minor (?) proportions of mixed layered illite-smectites andsmectites do occur (probably in part, the mixed layers represent a weathering product from the illite). Kaolinite is a variable component, but is not ubiquitous to all carbonate insoluble residues.

Calcrete associated insoluble residues are, however, dominated by mixed layer illite-smectites and smectites (best seen in tubulic sediment residues).Laminated calcrete crusts have a kaolinite-rich assemblage, most similar to that of suprapalaeokarstic wayboard clays. Further data may elucidate whether this clay mineral distribution is an effect of the contemporary weathering profiles, associated with the emergence events.

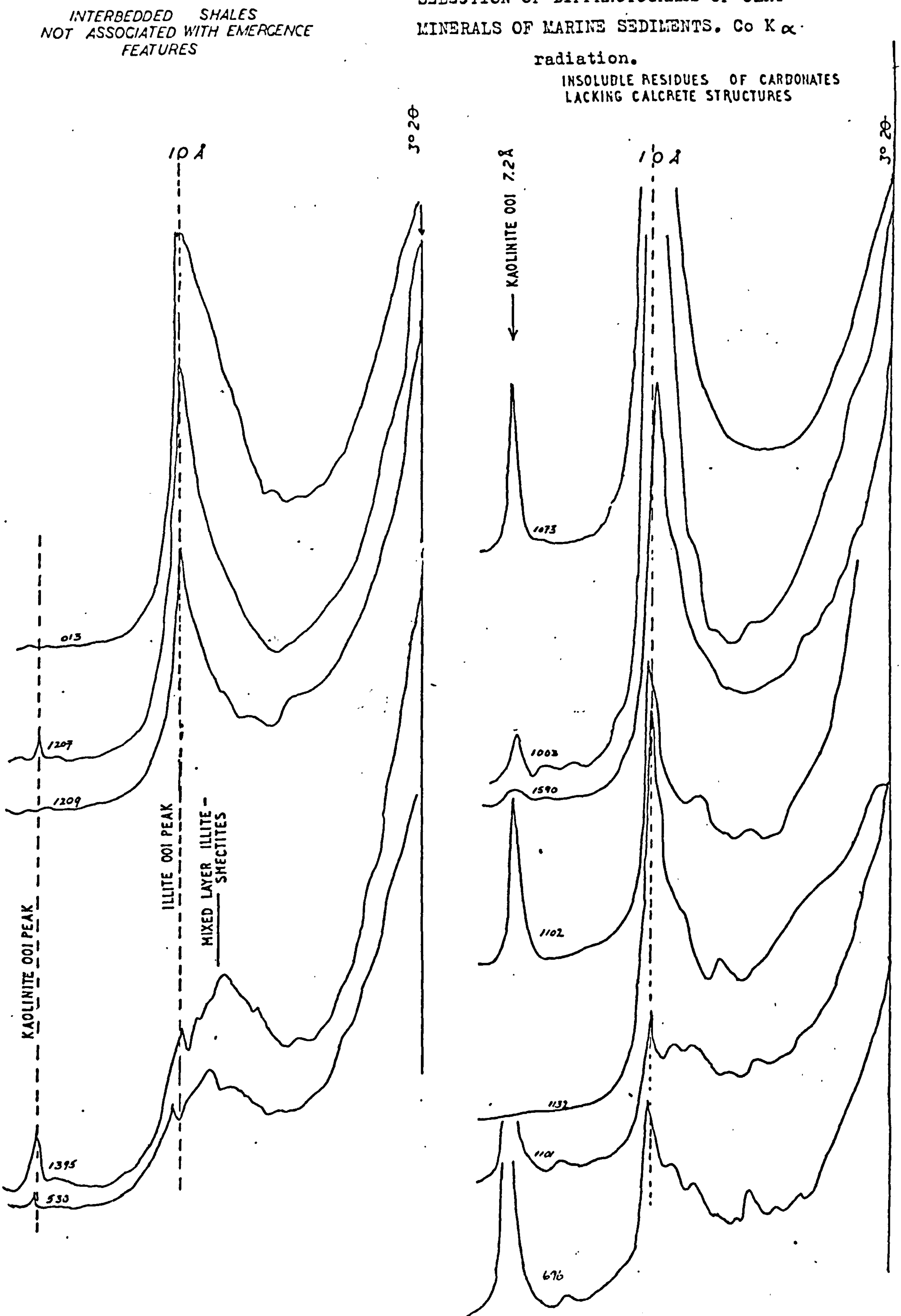
TABLE J

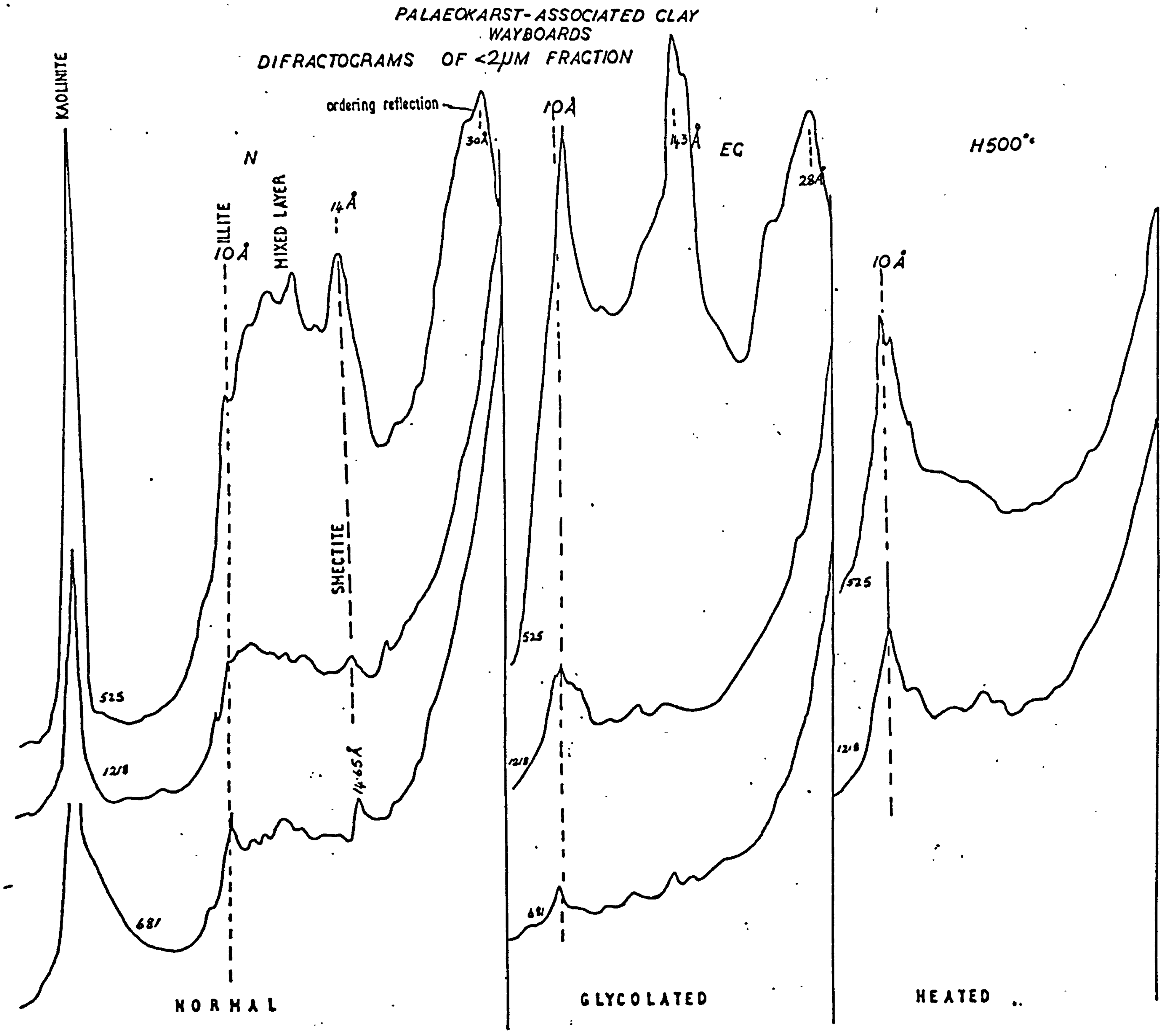
TABLE OF PEAK AREAS OF 001 REFLECTIONS EXPRESSED AS A PERCENTAGE OF THEIR TOTAL PEAK AREAS.

Description	Area % of 001 peak reflections of clays:					Lithology	Locality
	S/m N°	%Illite	%Illite-Smectite	%Smectite	%Kaolinite		
Pedotubule hor.ⁿ insol. residue	621	41	39	10	10	MA.1	Trefor Form'n,Trefor Rocks
	1280	29	50	11	11	MA.3	" ,Hafod,Minera
	1310	40	43	9	9	MA.1	Eglwyseg Form'n,Dinbren Uchaf
	757	19	43	14	24	MA.3	" ,Trefor Rocks
	1029	24	45	6	24	MA.3	" ,Bron Heulog
	1171	22	25	6	47	MA.3	" ,Trefor Rocks
Laminated crust insol. residue	1362	20	30	10	40		Trefor Fm'n,Llawnt
	621	16	28	8	48		" ,Trefor Rocks
	1504	40	15	5	40		Eglwyseg Fm'n, Tynant
	1027	22	37	4	37		" ,Bron Heulog
	1174	17	8	8	67		" ,Trefor Rocks
Wayboard clays – palaeokarst associated.	525	28	40	13	19		Brigantian, Corwen
	1218	25	38	8	30		Sychtyn M'br,Whitehaven
	681	24	24	6	47		Eglwyseg Fm'n,Bron Heulog
	530	31	58	8	4		Brigantian,Corwen
	1326	18	18	10	54		Sychtyn Mbr,Craig Sychtyn
insoluble residues	1073	92	0	0	8	MA.1	Sychtyn Mbr,Nant Mawr
	1003	91	3	1	5	MA.1	Tynant Fmn,Llwyn Hen Parc
	1590	92	4	0	4	MA.1	" ,Froncysyllte
	1102	64	11	4	22	MA.1	Sychtyn Mbr,Nant Mawr
	1132	86	9	6	0	MA.1	Tynant Fmn,Worlds End
	1101	34	23	9	34	MA.1	Sychtyn Mbr, Nant Mawr
	696	47	17	7	29	MA.3	Eglwyseg Fmn,Bron Heulog
Interbedded shales-non palaeokarst associated.	013	97	3	0	0		Tynant Fm'n , Tynant
	1207	97	0	0	3		'Basal Shales',Whitehaven
	1208	60	33	17	0		2" " "
	1209	93	5	2	0		" " "
	1210	77	8	8	8		" " "
	1395	34	53	2	11		Trefor Form'n,Dolgoch.

Fig A i

SELECTION OF DIFFRACTOGRAMS OF CLAY MINERALS OF MARINE SEDIMENTS. Co K α radiation.





DIFFRACTOGRAMS OF SUPRAPALAEOKARST CLAY WAYBOARDS.

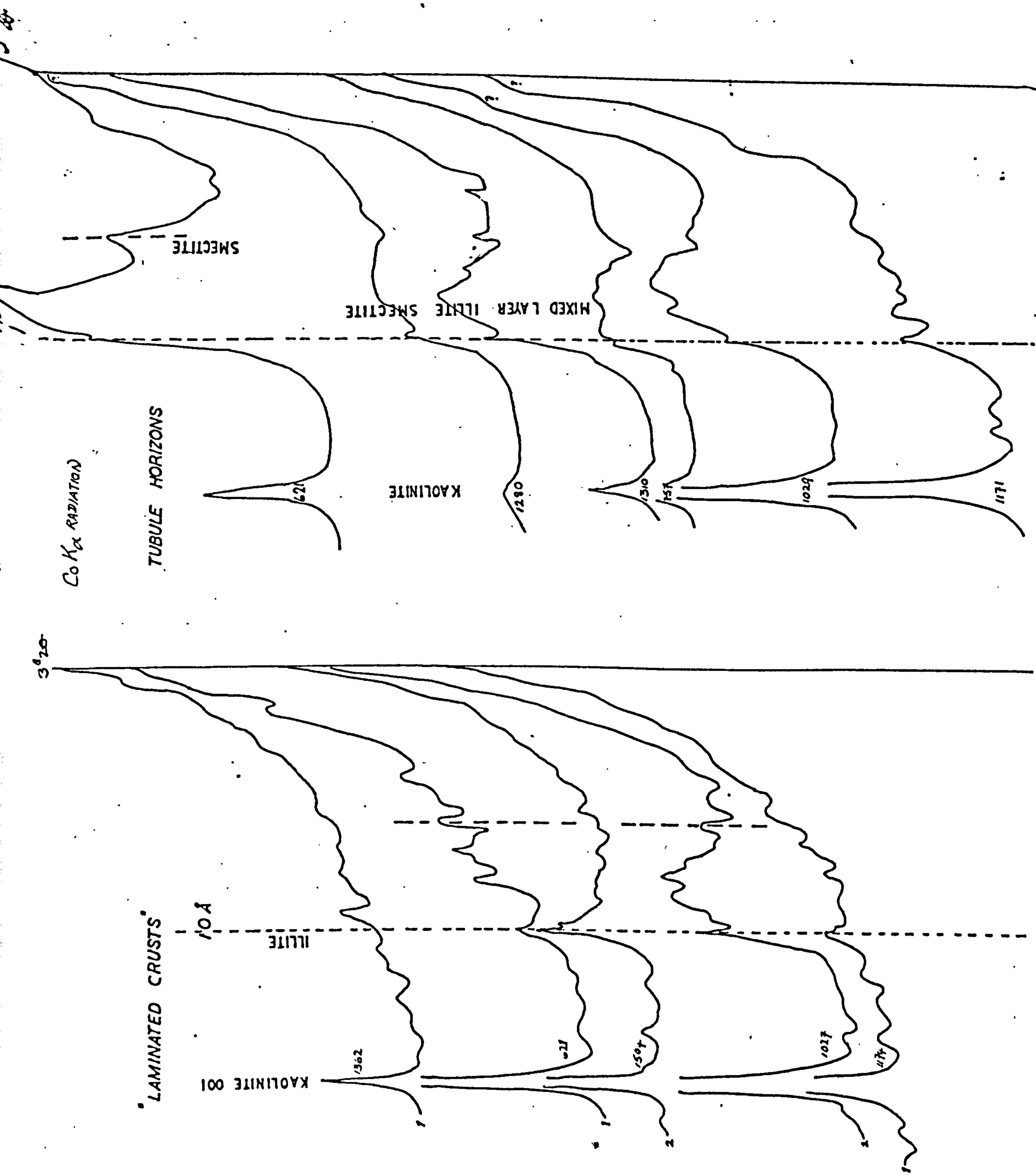
Co Kα radiation.

Fig Aii

Fig A iii

DIFFRACTOGRAMS OF CALCRETE CRUST AND TUBULAR CALCRETE

STRUCTURE HORIZONS - INSOLUBLE RESIDUES.



APPENDIX IV

List of fossils identified from Asbian sediments of the Llangollen and Oswestry embayments, Whilst providing a representative assemblage, this list is not exhaustive, and is based upon the authors own observations.

TYNANT FORMATION

Algae	?Coelosporella sp. Tynant, s/m 289.
	?Epistacheoides sp.~World's End, s/m 195.
	Garwoodia sp. nr Worlds End, s/m193
	Girvanella cf staminea Garwood Tynant, s/m 016.
	Kamaena cf delicata Antropov. Worlds End, s/m 163.
	Kamaenella denbighi Mamet & Roux. Throughout.
	Koninckopora inflata (de Koninck) Throughout
	Ortonella cf kershopensis Garwood. World's Ends/m192
	Pachysphaera spp Throughout
	Palaeocancellus spp Throughout
	Parathurammina cf suleimanovi Lipina
	Polyderma spp Throughout.
	Quasipolyderma sp.
	Pseudostacheoides sp.
	Ungdarella . Tynant, s/m004.
	Ungdarella sp.
	?Uraloporella sp Throughout.
	Vicinesphaera sp. Worlds End, s/m 163.
Foraminifera	Ammodiscids indet. Throughout
	Archaediscids indet. Throughout.
	Endothyranopsis sp. World's End, s/m 163
	Eostaffella sp Tynant, s/m238

	?*Palaeotextularia* sp. Tynant, s/m 142
	Plectogyra spp Throughout
Porifera	*Chaetetes* (*Boswellia*) *mortoni* sp.nov. Tynant, s/m248.
	Chaetetipora sp .Minera, s/m 092
Cnidaria	?*Cladochonus* sp. Tynant, s/m 289.
	Lithostrotion aranea (McCoy) Aber Sychnant
	L. cf *sociale* (Phillips)
	Syringopora sp. Tynant, s/m 026
Brachiopoda	*Camarotoechia* sp. Minera, s/m 1535
	Composita sp. Worlds End, s/m177
	Daviesiella llangollensis (Davidson) Throughout
	Linoprotonia cf *corrugatohemisphaerica* (Vaughan) Tynant, s/m
	L. hemisphaerica (J.Sowerby) group . Tynant, s/m 023.
Bryozoa	Fenestellid indet.nr Worlds End, s/m 193
Mollusca	Gastropods indet, (throughout, in section)
	Bivalves indet , (rare, in section only)
Arthropoda	Ostracods indet.
Echinodermata	Crinoids fragments , Throughout
	Echinoid spines. Throughout
Trace Fossils	*Zoophycos* sp. Tynant, s/m470

SYCHTYN MEMBER

Algae	Asphaltina sp.
	Calcispheres indet.
	Kamaena sp.
	Kamaenella cf denbighi Mamet & Roux.
	Koninckopora inflata (de Kononck)
	Ungdarella sp.
	?Uraloporella sp.
Foraminifera	Ammodiscids indet.
	Archaediscids indet.
	Endothyranopsis sp. s/m 1318
	Plectogyra spp
Porifera	Chaetetes (Boswellia) sp.
	C. cf tumidus von Waldheim.
Cnidaria	
	Lithostrotion cf martini Edwards & Haime
	Syringopora reticulata Goldfuss.
	S.sp.
Brachiopoda	Composita cf ficoides
	Daviesiella llangollensis Davidson.
	Gigantoproductus cf maximus (McCoy) group.
	Linoprotonia hemisphaerica (J.Sowerby) group.
Mollusca	Bivalves indet (in section only)
	Gastropods indet (in section only)
Arthropoda	Ostracods indet.
Echinodermata	Crinoid fragments & Echinoid spines.

L L A N Y M Y N E C H   M E M B E R

	Asphaltina sp.
Algae	Calcispheres indet.
	Kamaena sp.
	Kamaenella sp.
	Koninckopora sp.
	Ungdarella sp.
	Ammodiscids indet.
Foraminifera	Archaediscids indet.
	Endothyranopsis cf crassa (Brady)
	Eostaffella cf proikensis Rauser-Cernoussova s/m 1163
	?Palaeotextularia sp. s/m 1241
	Plectogyra spp throughout
	?Pseudoendothyra sp. s/m 1409
Porifera	Chaetetes (Boswellia) sp.
Cnidaria	Caninia sp.
	Dibunophyllum sp.
	Lithostrotion cf decipiens (McCoy)
	L. cf martini Edwards & Haime.
	Syringopora cf reticulata Goldfuss.
Brachiopoda	Composita sp.
	?Daviesiella sp.
	Linoprotonia hemisphaerica (J.Sowerby) group
Bryozoa	Cryptostome indet.
	Fenestella sp.
Mollusca	Bivalves indet (in section only)

	Gastropods indet (in section only)
<u>Arthropoda</u>	Ostracods indet.
<u>Echinodermata</u>	Crinoid fragments
	Echinoid spines.
<u>Trace fossils</u>	<u>Chondrites</u> sp.

E G L W Y S E G F O R M A T I O N (Below Eg.6)

Algae	Archaeosphaera sp. Minera, s/m 1538.
	Calcispheres indet Throughout,
	Kamaena delicata Antropov Throughout.
	Kamaena sp. Bron Heulog, s/m 268.
	Kamaenella denbighi Mamet & Roux Throughout.
	Koninckopora inflata (de Koninck) Throughout
	K. sp. (80um utricles) Minera, s/m 1184
	Pachysphaera cf polydermoides Conil & Lys. Minera, s/m1540
	P. cf dervillei Conil & Lys. Minera, s/m 1540.
	Ungdarella deceanglorum Elliot. Tynant, s/m 043
	U. sp. Minera, s/m 1577
	?Uraloporella sp. Throughout
	Vicinesphaera sp Minera, s/m 1540.
Foraminifera	Ammodiscids Throughout
	Archaediscids Throughout
	Cribrostomum sp. Minera, s/m 1538.
	Endothyranopsis crassa cf umbonata (Ganelina) Tan y Graig, s/m 106.
	Eostaffella sp.
	Plectogyra spp
Porifera	Chaetetes (Boswellia) sp.
	C. septosus (Fleming) Minera,
	Chaetetipora sp.
Cnidaria	Caninia cf juddi (Thomson) Minera, s/m 1291.
	Dibunophyllum bourtonense Garwood & Goodyear Trefor Rocks, s/m313.
	Hexaphyllia sp. Trefor Rocks, s/m 334.
	Koninckophyllum sp. Trefor Rocks, s/m 334c
	Lithostrotion cf decipiens (McCoy)
	L. aff irregulare Trefor Rocks, s/m1177.
	L. junceum (Fleming) Trefor Rocks, s/m 318.

	_L._cf _martini_ Edwards & Haime.
	_L._ _sociale_ (Phillips) Tynant,s/m 042.
	_L._ sp. nov. (J.R.Nudds pers. comm.) Bron Heulog, s/m260.
	Palaeosmilia _murchisoni_ (Edwards & Haime). Trefor Rocks,s/m 316.
	Syringopora cf _distans_ (Fischer)
	_S._ cf _gigantea_ Thomson
	_S._ _reticulata_ Goldfuss
	_S._ _ramulosa_ Goldfuss
Bryozoa	?_Stenopora_ sp. Craig Arthur,s/m 104
Brachiopoda	'_Athyris_' cf _planosulcata_ McCoy.Llwyn Hen Park, s/m1238.
	Brachythyris sp. Trefor Rocks,s/m763.
	Composita sp. Bron Heulog, s/m 253 .
	Daviesiella _llangollensis_ Davidson.Minera,s/m1534.
	Delepinea aff _comoides_ (J.Sowerby)
	Gigantoproductus _maximus_ (McCoy) group Trefor Rocks, 11 6.
	Linoprotonia _hemisphaerica_ (J.Sowerby) group. Throughout
	Megachonetes sp. _papilonaceous_ (Phillips) group Trefor Rocks,s/m314.
	Plicochonetes sp. Trefor Rocks,s/m315
	Rugosochonetes _hardrensis_ (Phillips)Trefor Rocks, s/m 763.
Mollusca	Bivalves indet Trefor Rocks, s/m 1177.
	Gastropods indet Throughout
	Orthocone indet. Minera, 1185.
	Straparollus (_Euomphallus_) sp. Llwyn Hen Parc,s/m 1238.
Arthropoda	Ostracods indet.
Echinodermata	Crinoid & Echinoid fragments.

E G L W Y S E G F O R M A T I O N (Above'Eg. 6')

Calcispheres indet Throughout

Algae

Kamaena delicata Antropov Throughout.

Kamaena sp. Tynant, s/m 049

Kamaenella denbighi Mamet & Roux Tynant ,s/m 053

Koninckopora spp Throughout

Ungdarella deceanglorum Elliot. Bron Heulog,s/m 255

U. sp. e.g. Bron Heulog, s/m 260.

?*Uraloporella* sp. Throughout.

Foraminifera

Ammodiscids Throughout

Archaediscids Throughout

Cribrostomum sp Tynant 048

Endothyranopsis spp Throughout

Eotuberitina sp. Froncysyllte,s/m1593.

Plectogyra spp Throughout

Porifera

Chaetetes septosus (Fleming) Bron Heulog,s/m 119

C. cf *tumidus* von Waldheim Bron Heulog,s/m 2548

Cnidaria

Dibunophyllum aff *bourtonense* Garwood & Goodyear Trefor Rocks,748

D. sp. Trefor Rocks,s/m 748b

Hexaphyllia sp. Trefor Rocks, s/m 763

Lithostrotion ?*irregulare* (Phillips) 691,Bron Heulog.

L. junceum (Fleming) Trefor Rocks,s/m 750.

L. aff *junceum*(Fleming) Trefor Rocks, s/m 740 (large form)

L. maccoyanum Edwards & Haime. T ynant,s/m 304.

L. cf *martini* Edwards & Haime Bron Heulog, s/m 691

L. aff *pauciradiale* (McCoy) Trefor Rocks,s/m759

Palaeosmilia murchisoni (Edwards & Haime) Bron Heulog,s/m693

?*Siphonophyllia* sp Tynant, s/m 46.

	Syringopora reticulata Goldfuss. Bron Heulog, s/m 1024.
Brachiopoda	*Brachythyris* sp. Froncysyllte, s/m1592
	Composita sp. Froncysyllte, s/m1592.
	Gigantoproductus maximus(McCoy) group Tan y Graig, s/m 1273.
	G. sp. Llwyn Hen Parc, s/m1250.
	Linoprotonia hemisphaerica (J.Sowerby) group. Throughout
Bryozoa	Fenestellid indet Froncysyllte, 1593
	Trepostome indet (encrusting form)
Mollusca	Gastropods indet Throughout
	Edmondia sp. Tynant, s/m1228.
Arthropoda	Ostracods indet Throughout
Echinodermata	Crinoid fragments Throughout
	Echinoid spines Throughout.

LLYNCLYS FORMATION

Algae	Calcispheres indet.
	Kamaena sp.
	Kamaenella denbighi Mamet & Roux
	Koninckopora cf inflata (de Koninck)
	Ungdarella sp
	Ammodiscids indet
Foraminifera	Archaediscids indet.
	Cribrostomum sp.
	Endothyranopsis crassa subsp.(Brady)
	Eostaffella sp.
	Plectogyra spp .
Porifera	Chaetetes septosus (Fleming)
Cnidaria	Dibunophyllum bourtonense Garwood & Goodyear
	Hexaphyllia sp.
	Lithostrotion aff irregulare (Phillips)
	L. martini Edwards & Haime
	L. sociale (Phillips)
	Syringopora cf distans (Fischer)
	Syringopora cf reticulata Goldfuss
Brachiopoda	Gigantoproductus cf maximus (McCoy) group
	Linoprotonia hemisphaerica (J. Sowerby) group.
	Rugosochonetes sp.
Mollusca	Gastropods indet
	?Naticopsis sp.

Arthropoda	Ostracods indet.
	Trilobite fragments.
Echinodermata	Crinoid fragments
	Echinoid spines

APPENDIX V

List of fossils identified from sediments below Tf.7, Trefor Formation, between Minera and Dolgoch. (Lower Brigantian).

Algae

?Asphaltina sp. Craignant,s/m 722

Bisphaera sp. Craignant,s/m722

Coelosporella jonesi Wood , throughout.

Coelosporella sp. Bron y Garth s/m 521 (c.80µm utricles)

Girvanella sp. Craignant, s/m722

?Kamaenella sp. Bron y Garth,s/m 561.

Kamaena sp. Trefor Rocks,s/m209

Nanopora cf anglica Wood, Froncysyllte,s/m426.

Pachysphaera sp. throughout

Ungdarella spp. Plas Ifa,1453;Froncysyllte,426

(Calcispheres indet. throughout)

Foraminifera

Ammodiscids indet. throughout.

Archaediscids throughout.

Earlandia sp. Plas Ifa s/m 433

Endothyranopsis crassa umbonata (Ganelina), Tynant, 071

?Loeblichia sp. Tynant ,s/m 079

?Millerella sp. Tynant,s/m 227

Palaeotextularia sp. Tynant, s/m199

Plectogyra spp throughout

Saccamminopsis sp. Plas Ifa,432.

Textularids indet throughout

Tetrataxis cf conica Ehrenberg Tynant, 067.

Valvulinella sp.Trefor .Rocks,s/m200.

Porifera

Chaetetes (Boswellia) sp. Trefor Rocks,s/m 070

C.depressus Flem ing Treflach,s/m1059.

Chaetetes septosus Flemming. s/m 209, Trefor Rocks.

Cnidaria

Aulophyllum fungites (Fleming) Tf1.2.

Aulopora sp. Dolgoch, s/m 1384

?Cladochonus sp. Froncysyllte, s/m426.

Dibunophyllum bipartitum bipartitum (McCoy) Bron y Garth, 540

D.b. craigianum (Thomson) Froncysyllte, s/m 423.

D.sp. (lamellar torsion) Llawnt, s/m1366.

Diphyphyllum cf fasciculatum (Flemming) Craignant, s/m733

D. furcatum (Thomson) Pant Hir, s/m1197.

D. cf gracile Bron y Garth, s/m543.

Hexaphyllia sp. Froncysyllte, 426.

Lithostrotion decipiens (McCoy) Bron y Garth, s/m 540.

L. junceum (Flemming) Bron y Garth, 544

L. irregulare , Dolgoch, 1383.

L. cf maccoyanum Milne Edwards & Haime, Pant Hir, s/m1336.

L. martini Milne Edwards & Haime, Bron y Garth s/m 556.

L. pauciradiale (McCoy) Dolgoch, s/m 1383.

Lonsdaleia floriformis floriformis (Martin) - small form of Smith (1930), Pant Hir, s/m1335

L. f. subsp. Pant Hir, 2012.

Palaeosmilia regia (Phillips) Tynant, 79.

Siphonophyllia sp. Froncysyllte, s/m 424.

Syringopora spp Throughout.

Bryozoa

Trepostome indet. Tynant, s/m067

Fenestella sp Plas Ifa s/m431 & throughout

?Rhabdomeson sp. Froncysyllte, s/m424.

Brachiopoda

Athyris planosulcata Dolgoch, 1388

Brachythyris sp. Treflach, s/m1060.

Cleiothyris roysii Treflach, 1060

Composita ambigua Throughout

Gigantella cf crassiventer Prentice Dolgoch,1388

Gigantella sp. Minera,1281

Gigantoproductus sp maximus(McCoy)group Dolgoch,1394

Gigantoproductus cf okensis Dolgoch,s/m1386

G. sp. Craig y Rhiw,s/m1372.

?Leptagonia sp. Dolgoch,s/m1393

Linoprotonia sp. Minera.

Megachonetes sp. Trefor Rocks,Tf1.2a.

Overtonia fimbriata (J de C Sowerby) Pant Hir,s/m2009.

Productus (Dictyoclostus) pinguis (Muir-Wood)Trefor Rocks s/m 558.

Productus productus (Matin) Trefor Rocks,071.

Semiplanus sp latissimus (J.Sowerby) group Throughout.

Schellwienella sp . Trefor Rocks,Tf1.2a.

?Reticularia sp. Treflach,1060.

Bryozoa

Fenestella spp. Throughout.

?Rhabdomeson sp. Froncysyllte, s/m 421.

Trepostome indet. Tynant,s/m227.

Mollusca

Bivalves indet Throughout.

?Edmondia sp, Tynant,s/m 702.

Gastropods indet, common throughout.

Arthropoda

?Weberides sp. Trefor Rocks, s/m 198.

Ostracods indet. throughout.

Echinodermata

Crinoid & Echinoid fragments throughout.

List of fossils identified from the Localities, unless otherwise stated are along the scarp of Morton's Coral Bed between Tynant and Eglwyseg Plantation. Macrofauna only.

Porifera	Chaetetes depressus (Fleming)
	C. septosus (Fleming)
Cnidaria	Aulopora sp.
	Caninia cf juddi (Thomson)
	?Cladochonus sp.
	Clisiophyllum keyserlingi McCoy.
	Corwenia rugosa (McCoy)
	Dibunophyllum bipartitum bipartitum(McCoy)
	D.b.konincki (Edwards & Haime)
	D.b. craigianum (Thomson)
	Diphyphyllum cf lateseptatum McCoy.
	Koninckophyllum cf magnificum Thomson & Nicholson
	Lithostrotion decipiens (McCoy)
	L. junceum (Fleming)
	L.aff maccoyanum Edwards & Haime
	L. martini Edwards& Haime
	L. pauciradiale (McCoy)
	Lonsdaleia duplicata duplicata(Martin)
	L.d. subsp. (with gastrovascular interconnection)
	L. floriformis floriformis (Martin) small form Smith 1916)
	Nemistium edmondsi Smith
	Orionastraea tuberosa (Phillips)
	Palaeosmilia regia (Phillips)
	Syringopora catenata (Martin)
	S.cf reticulata Goldfuss
	?Zaphrentites sp.

Brachiopoda

'Athyris' cf planosulcata var obtusa McCoy.

Athyris cf expansa (Phillips)

Antiquatonia sp Corwen

Avonia sp.

Brachythyris sp.

Buxtonia aff scabricula (Martin)

'Camarotoechia' pleurodon (Phillips)

Cleiothyris roysii (Levillier)

Composita ambigua (J.Sowerby)

Dielasma sp.

Gigantella sp.

Gigantoproductus sp. maximus(McCoy) group

G.cf okensis ; Corwen.

G. aff striatosulcatus

G.sp.

?Latiproductus sp.

Martinia glabra Martin.

Overtonia fimbriata (J.de C.Sowerby)

Phricodothyris sp.

Plicochonetes sp,

Productus (Dictyoclostus) aff insculptus Muir-Wood

P. (D.) aff muiricatus (Phillips)

P. (D.) sulcatus (J.Sowerby)

P.(Echinoconchus) punctatus (J.Sowerby)

P. (Eomarginifera) aff derbiensis (Muir-Wood)

P. aff mesolobatus Phillips Corwen.

P. productus Martin

P.(Pustula) elegans

Ptilothyris laminosa)(J.Sowerby)

?Reticularia sp. Corwen,s/m 514

Rugosochonetes cf hardrensis

	Rugosochonetes sp.
	Schizophoria aff *resupinata* (Martin)
	Semiplanus fragilis
	Semiplanus sp.
	Spirifera cf *bisulcata* J de C Sowerby, transverse form.
	?*Syringothyris* sp.
Bryozoa	*Fenestella* spp
	Polypora sp.
	Stenopora sp.
	Trepostome indet.
Mollusca	*Aviculopecten* sp
	Euphemites cf urii (Fleming)
	?*Palaeozygopleura* sp.
	Pinna sp.
	Sanguinolites sp.
Echinodermata	Crinoid fragments
	Echinoid spines.

APPENDIX VI

Some Asbian Calcareous Algae (Plates 23 and 24)

PLATE 23

FIG A. . . ?*Stacheoides* sp. ,s/m 1146, Llanymynech Member.

FIG B.& D. : *Pseudostacheoides* sp. Branching, cellular thalli.

B. s/m o31, Eglwyseg Formation ,Tynant.

D. s/m 195, upper cycle of Tynant Formation, 1km N of World's End.

(see Petryk & Mamet,1972)

FIG C. *Ungdarella* sp. s/m 004, Tynant Formation, Tynant.

FIG.E. *Ungdarella deceanglorum* Elliot. s/m 255, Eglwyseg Form'n

PLATE 23
(APPENDIX)

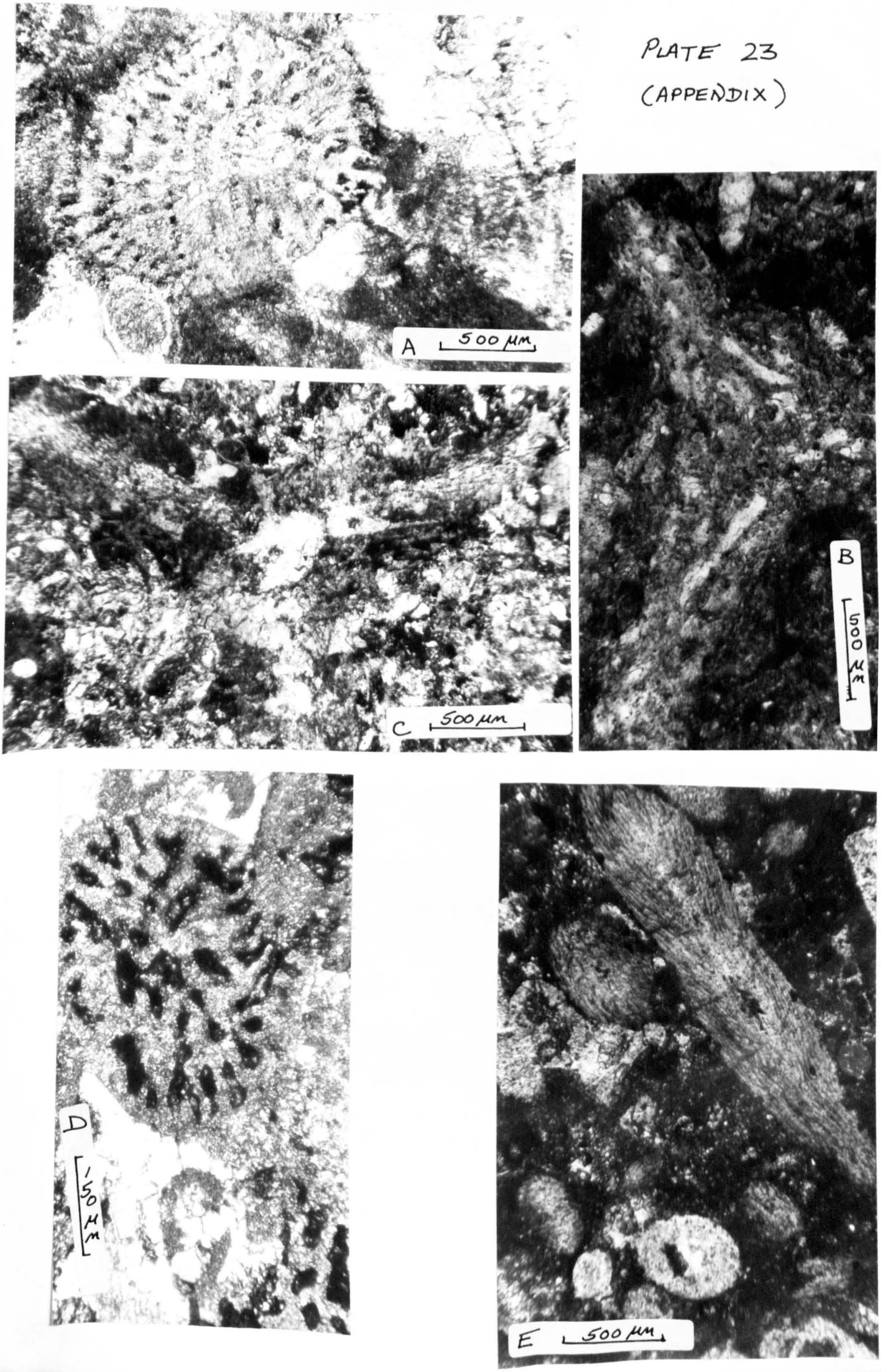


PLATE 24

FIG A. ?*Asphaltina* sp. s/m 742, Eg5.6, . Trefor Rocks.

FIG B. Grainstone - packstone laminae rich in beresellids, vaguely discernible by inclusion rich walls in syntaxial spar. s/m 049, Eglwyseg Formation, Tynant. ?*Kamaenella* sp. (*Uraloporella* may be distinguished by its complete septal partitions to the medula.)

FIG C. *Kamaenella denbighi* (Mamet & Roux. Note incomplete nature of septal partitions. s/m 004, Tynant Formation, Tynant.

FIG. D. *Kamaena* cf *delicata* Antropov. s/m 255, Eglwyseg Formation, Bron Heulog.

A

B

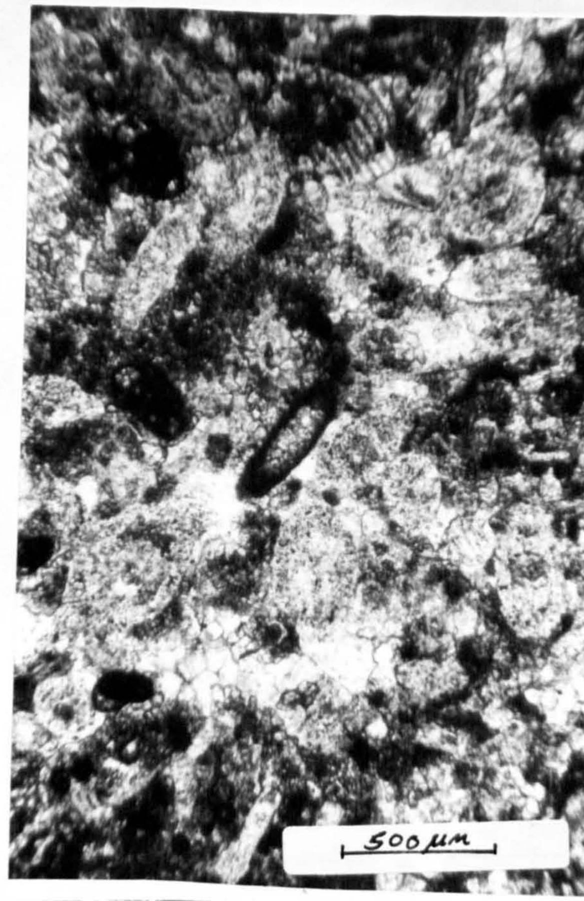
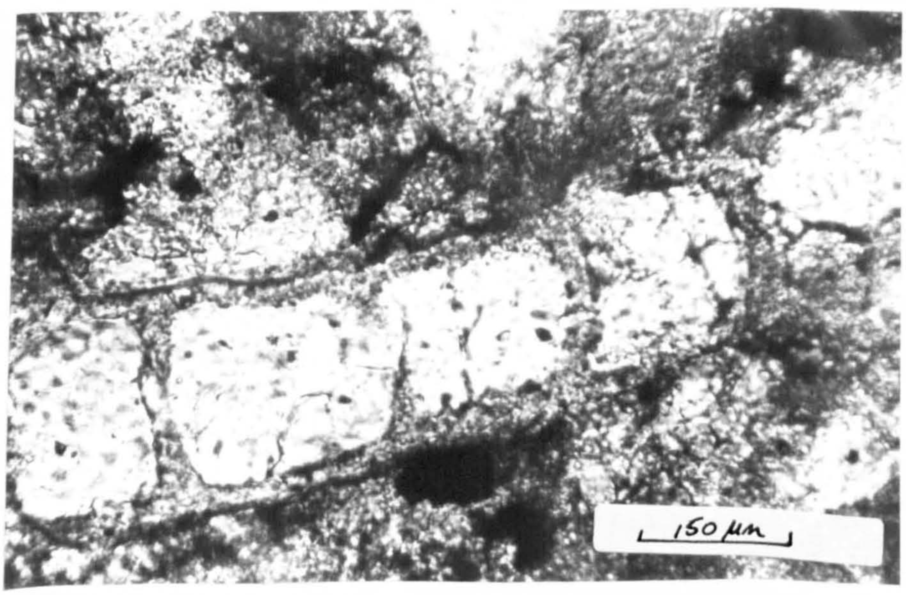


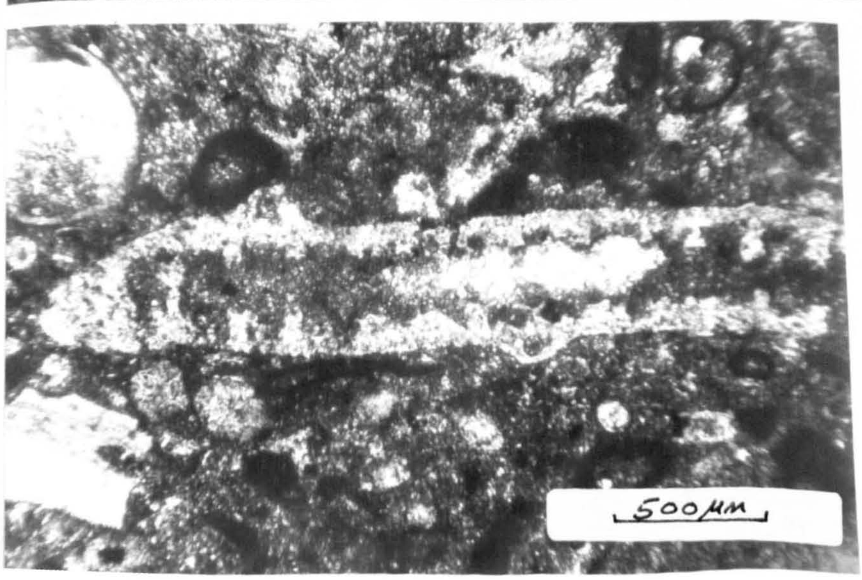
PLATE 24

(APPENDIX)



C

D



Reprinted from PALAEONTOLOGY *Volume* 23 *Part* 4
November 1980

Spicule pseudomorphs in a new Palaeozoic chaetetid and its sclerosponge affinities

BY
DAVID I. GRAY

PUBLISHED BY THE
PALAEONTOLOGICAL ASSOCIATION
LONDON

THE PALAEONTOLOGICAL ASSOCIATION

The Association publishes *Palaeontology* and *Special Papers in Palaeontology*. Details of membership and subscription rates may be found inside the back cover.

The journal *Palaeontology* is devoted to the publication of papers on *all aspects* of palaeontology. Review articles are particularly welcome, and short papers can often be published rapidly. A high standard of illustration is a feature of the journal. Four parts are published each year and are sent free to all members of the Association. *Typescripts* on all aspects of palaeontology and stratigraphical palaeontology are invited. They should conform in style to those already published in this journal, and should be sent to Dr. R. A. Fortey, Palaeontological Association, Department of Palaeontology, British Museum (Natural History), Cromwell Road, London SW7 5BD, England, who will supply detailed instructions for authors on request (these were published in *Palaeontology* 1977, **20,** pp. 921–929).

Special Papers in Palaeontology is a series of substantial separate works; the following are available (post free).

1. (for 1967): Miospores in the Coal Seams of the Carboniferous of Great Britain, *by* A. H. V. SMITH *and* M. A. BUTTERWORTH. 324 *pp.*, 72 *text-figs.*, 27 *plates.* Price £8 (U.S. $19.50).
2. (for 1968): Evolution of the Shell Structure of Articulate Brachiopods, *by* A. WILLIAMS. 55 *pp.*, 27 *text-figs.*, 24 *plates.* Price £5 (U.S. $12).
3. (for 1968): Upper Maestrichtian Radiolaria of California, *by* HELEN P. FOREMAN. 82 *pp.*, 8 *plates.* Price £3 (U.S. $7.50).
4. (for 1969): Lower Turonian Ammonites from Israel, *by* R. FREUND *and* M. RAAB. 83 *pp.*, 15 *text-figs.*, 10 *plates.* Price £3 (U.S. $7.50).
5. (for 1969): Chitinozoa from the Ordovician Viola and Fernvale Limestones of the Arbuckle Mountains, Oklahoma, *by* W. A. M. JENKINS. 44 *pp.*, 10 *text-figs.*, 9 *plates.* Price £2 (U.S. $5).
6. (for 1969): Ammonoidea from the Mata Series (Santonian-Maastrichtian) of New Zealand, *by* R. A. HENDERSON. 82 *pp.*, 13 *text-figs.*, 15 *plates.* Price £3 (U.S. $7.50).
7. (for 1970): Shell Structure of the Craniacea and other Calcareous Inarticulate Brachiopoda, *by* A. WILLIAMS *and* A. D. WRIGHT. 51 *pp.*, 17 *text-figs.*, 15 *plates.* Price £1·50 (U.S. $4).
8. (for 1970): Cenomanian Ammonites from Southern England, *by* W. J. KENNEDY. 272 *pp.*, 5 *tables*, 64 *plates.* Price £8 (U.S. $19.50).
9. (for 1971): Fish from the Freshwater Lower Cretaceous of Victoria, Australia, with Comments on the Palaeo-environment, *by* M. WALDMAN. 130 *pp.*, 37 *text-figs.*, 18 *plates.* Price £5 (U.S. $12).
10. (for 1971): Upper Cretaceous Ostracoda from the Carnarvon Basin, Western Australia, *by* R. H. BATE. 148 *pp.*, 43 *text-figs.*, 27 *plates.* Price £5 (U.S. $12).
11. (for 1972): Stromatolites and the Biostratigraphy of the Australian Precambrian and Cambrian, *by* M. R. WALTER. 268 *pp.*, 55 *text-figs.*, 34 *plates.* Price £10 (U.S. $24).
12. (for 1973): Organisms and Continents through Time. A Symposium of 23 papers edited by N. F. HUGHES. 340 *pp.*, 132 text-figs. Price £10 (U.S. $24) (published with the Systematics Association).
13. (for 1974): Graptolite studies in honour of O. M. B. Bulman. Edited by R. B. RICKARDS, D. E. JACKSON, *and* C. P. HUGHES. 261 *pp.*, 26 *plates.* Price £10 (U.S. $24).
14. (for 1974): Palaeogene Foraminiferida and Palaeoecology, Hampshire and Paris Basins and the English Channel, *by* J. W. MURRAY *and* C. A. WRIGHT. 171 *pp.*, 45 *text-figs.*, 20 *plates.* Price £8 (U.S. $19.50).
15. (for 1975): Lower and Middle Devonian Conodonts from the Broken River Embayment, North Queensland, Australia, *by* P. G. TELFORD. 100 *pp.*, 9 *text-figs.*, 16 *plates.* Price £5·50 (U.S. $13.50).
16. (for 1975): The Ostracod Fauna from the Santonian Chalk (Upper Cretaceous) of Gingin, Western Australia, *by* J. W. NEALE. 131 *pp.*, 40 *text-figs.*, 22 *plates.* Price £6·50 (U.S. $16).
17. (for 1976): Aspects of Ammonite Biology, Biogeography, and Biostratigraphy, *by* W. J. KENNEDY *and* W. A. COBBAN. 94 *pp.*, 24 *text-figs.*, 11 *plates.* Price £6 (U.S. $14.50).
18. (for 1976): Ostracoderm Faunas of the Delorme and Associated Siluro-Devonian Formations, North West Territories, Canada, *by* D. L. DINELEY *and* E. J. LOEFFLER. 218 *pp.*, 78 *text-figs.*, 33 *plates.* Price £20 (U.S. $48).
19. (for 1977): The Palynology of Early Tertiary Sediments, Ninetyeast Ridge, Indian Ocean, *by* E. M. KEMP *and* W. K. HARRIS. 74 *pp.*, 2 *text-figs.*, 8 *plates.* Price £7 (U.S. $17).
20. (for 1977): Fossil Priapulid Worms, *by* S. C. MORRIS. 159 *pp.*, 99 *text-figs.*, 30 *plates.* Price £16 (U.S. $38.50).
21. (for 1978): Devonian Ammonoids from the Appalachians and their bearing on International Zonation and Correlation, *by* M. R. HOUSE. 70 *pp.*, 12 *text-figs.*, 10 *plates.* Price £12 (U.S. $29).
22. (for 1978, published 1979): Curation of palaeontological collections. A joint colloquium of The Palaeontological Association and Geological Curators' Group. Edited by M. G. BASSETT. 280 *pp.*, 53 *text-figs.* Price £25 (U.S. $60).
23. (for 1979): The Devonian System. A Palaeontological Association International Symposium. Edited by M. R. HOUSE, C. T. SCRUTTON, *and* M. G. BASSETT. 353 *pp.*, 102 *text-figs.*, 1 *plate.* Price £30 (U.S. $72).
24. (for 1980): Dinoflagellate Cysts and Acritarchs from the Eocene of southern England, *by* J. B. BUJAK, C. DOWNIE, G. L. EATON, *and* G. L. WILLIAMS. 104 *pp.*, 24 *text-figs.*, 22 *plates.* Price £15 (U.S. $36).

SPICULE PSEUDOMORPHS IN A NEW PALAEOZOIC CHAETETID, AND ITS SCLEROSPONGE AFFINITIES

by DAVID I. GRAY

ABSTRACT. A Palaeozoic chaetetid, bearing intramural spicule pseudomorphs, *Chaetetes* (*Boswellia*) *mortoni* sp. nov., is described from the British Dinantian. Spicules are preserved as calcite, pyrite, and silica pseudomorphs. Only silica pseudomorphs retain detail of their tylostyle form. Neomorphism locally obliterates the spicular fabric. A primary mineralogy is suggested consisting of an aragonitic calcareous skeleton, with entrapped opal 'A' spicules. Comparison of morphology and microstructure with extant and fossil sclerosponges indicates a close relationship between this chaetetid and the Ceratoporellida, and supports the sclerosponge nature of some Palaeozoic chaetetids.

THE Class Sclerospongiae Hartman and Goreau, 1972, was proposed following the rediscovery of coralline sponges among the Jamaican coral-reef ahermatypic cryptofauna (Hartman 1969; Hartman and Goreau 1970). Sclerosponges were defined by Hartman and Goreau (1972, p. 144) as 'sponges secreting a compound skeleton of siliceous spicules, proteinaceous fibres and calcium carbonate, the latter laid down as a basal mass in which the siliceous spicules may or may not be entrapped'. The similarity of fossil chaetetids to some sclerosponges (briefly discussed by Kirkpatrick (1909, 1912*a*, 1912*b*) along with the monticuliporans), led Hartman and Goreau (1972) to remove the Chaetetida Okulitch, 1936, from the Anthozoa or Hydrozoa to the Sclerospongiae. They also erected the Order Ceratoporellida Hartman and Goreau, 1972, to include four extant sclerosponge genera, and added a third Order, the Tabulospongida Hartman and Goreau, 1975, following the discovery of a tabula-bearing form from the Pacific (Hartman and Goreau 1975), of which two more extant species have subsequently been described (Mori 1976, 1977) and a record traced back into the Mesozoic. Stearn (1972, 1975) discussed the sclerosponge affinities of the stromatoporoids.

The recognition of the Sclerospongiae as a Class has been questioned by a number of authors. Lévi (1973) considered the sclerosponges as a Subclass of the Demospongiae, subsequently followed by Vacelet, Vasseur, and Levi (1976) and Vacelet (1977). This classification takes into account the organization of living sclerosponge tissue which is 'basically similar to that of the Class Demospongiae except that it is divided into units each of which extends down into the upper layer of the basal calcareous skeleton' (Hartman and Goreau 1972, pp. 144–145).

The variability in spicule form and distribution (see Table 1) suggests that some sclerosponges may be related even more closely to other groups of demosponges. For example, Vacelet (1977, p. 347) mentions that the spicule character of *Tabulospongia wellsi* is similar to that displayed by the Spirastrellidae. Vacelet (1977, p. 347) also states that a basal calcareous skeleton may be a convergent structure in many groups of demosponges. This would account for the great variability of calcareous skeletal morphology and microstructure (see Table 1) observed in those forms classified as sclerosponges, and would imply that 'sclerosponge' is a convenience term for considering groups with a similar homeomorphic tendency.

In this paper the sclerosponges are considered as a Subclass of the Class Demospongiae Sollas, 1875.

The fossil history of sclerosponges that entrap spicules in their calcareous skeleton is represented by a limited assortment of forms including a few ceratoporellids, one species (Kaźmierczak 1974) of

[Palaeontology, Vol. 23, Part 4, 1980, pp. 803–820, pls. 102–103.]

the Order Muranida Kaźmierczak and Hillmer, 1974, a few problematical records of stromatoporoids and, until now, only two species of Mesozoic chaetetids. Table 1 summarizes their distribution and variation and allows comparisons to be made with extant forms.

Chaetetids are a diverse group with separate Palaeozoic and Mesozoic histories. Differences exist in the skeletal architecture of Palaeozoic and Mesozoic forms (Fischer 1970), and their phylogenetic relationships are not completely understood. Scrutton (1979, p. 169) reviewed briefly their relationships, and whilst supporting their sclerosponge affinities he emphasized the lack of convincing spicules associated with chaetetids as 'a major source of doubt for some workers' to their classification within the Porifera. Dieci, Russo, Russo, and Marchi (1977) were the first to report a spicule-bearing 'chaetetid', *Atrochaetetes medius* Cuif and Fischer, 1974, from the Upper Triassic of Italy, with intramural acanthostyle spicules, replaced by calcite. *Atrochaetetes* Cuif and Fischer, 1974, is characterized by a discontinuous backfill of fascicular fibrous carbonate extending into the lumen (Cuif and Fischer, 1974, p. 8) rather than complete tabulae typical of the chaetetids *s.s.* Continuous fascicular fibrous backfills are typical of ceratoporellids (see below). Since *A. medius* also has a ceratoporellid-like spicular fabric, the genus *Atrochaetetes* should be regarded as an aberrant member of the Ceratoporellida, and removed from the Chaetetida.

Kaźmierczak (1979) reported intramural monaxon spicules, replaced by pyrite, within a Lower Cretaceous (Barremian) chaetetid, *Chaetetopsis favrei* (Deninger 1906) from the Crimea. Like many Mesozoic chaetetids, *C. favrei* increases both by intramural offset and longitudinal (pseudoseptal) fission. The former is not known to occur in Palaeozoic chaetetids (Sokolov 1962, p. 262). The walls of *C. favrei* are anhedral calcite miscospar that is possibly a neomorphic overprint (Kaźmierczak 1979, p. 101), and is dissimilar from the typical fascicular fibrous microstructure of Palaeozoic chaetetids.

This is the first report of convincing intramural spicule pseudomorphs in a Palaeozoic chaetetid, *Chaetetes* (*Boswellia*) *mortoni* sp. nov., from the Lower Carboniferous of north Wales, northern England, and southern Scotland. Comparison is made with other chaetetids and sclerosponges, and a model is developed for the mode of spicule preservation. Classification of the Chaetetida within the Sclerospongidea is supported.

OCCURRENCE AND PRESERVATION OF MATERIAL

Eight colonies of this new species have recently been collected from the Lower Asbian (Upper Dinantian), Tynant Limestone (Somerville, 1979) (Lower Brown Limestone of Morton, 1879) of the Llangollen area, north Wales. Although this sclerosponge is a rare element in the brachiopod-dominated macrofauna, it has been collected from a 20-m range of cyclic strata at sites over 4 km of outcrop, and from the underlying scree. It occurs towards minor cycle bases, in subtidally deposited argillaceous algal-foraminiferal packstones and grainstones. The colonies were rolled and some were fragmented prior to burial. One colony has a pronounced micritic (?endolithic algal) rim on part of its

TABLE 1. Table of extant and fossil sclerosponges with associated spicules showing their spicule form and relationship to the basal calcareous skeleton. Mesozoic stromatoporoids of Schnorf (1960) and Yabe and Sugiyama (1935) are omitted owing to their uncertain spicular nature. Species of *Leiospongia* d'Orbigny, 1850, and *Hartmanina* Dieci *et al.* 1974, described by Dieci, Russo, and Russo (1974*b*) are ommitted owing to the absence of associated spicules.

Order symbols, Cer = Ceratoporellida, Tab = Tabulospongida, Unas = Unassigned, Mur = Muranida, Ch = Chaetetida, Stp = Stromatoporoidea; calcareous microstructure symbols, Fascic. fib. = fascicular fibrous, Agg. spher. = aggregated spherules, Microgran. = microgranular; original mineralogy symbols, A = aragonite, Mg-cc. = high magnesian-calcite; spicule distribution symbols, I. = intramural, E. = extramural, m. = subparallel to microstructure fibres, s. = subparallel to growth axis of skeleton, d. = embedded only within the distal portion of the calcareous skeleton, r. = random; spicule type symbols, * = megasclere, ** = microsclere; spicule mineralogy symbols, cc. = calcite, pyr. = pyrite, Fe ox. = iron oxide.

AGE	ORDER	SCLEROSPONGE	SOURCE OF DATA	CALCAREOUS SKELETON: MICRO-STRUCTURE	CALCAREOUS SKELETON: ORIGINAL MINERALOGY	SPICULE DATA: DISTRI-BUTION	SPICULE DATA: TYPE	SPICULE DATA: SIZE RANGE Length (μm)	SPICULE DATA: SIZE RANGE Diam.(μm)	SPICULE DATA: PRESENT MINERALOGY
EXTANT	Cer	Ceratoporella nicholsoni	Hartman & Goreau I970	Fascic.fib.	A	I.m.	Acanthostyle*	206 - 298	3.I - 4.0	Opal'A'
		Stromatospongia vermicola	"	"	"	"	"	I65 - I87	6.2 - 8.0	"
		Stromatospongia norae	"	"	"	"	"	I95 - 2I5	5.5 - 6.I	"
		Hispidopetra miniana	"	"	"	"	Style*	269 - 30I	5.4 - 7.4	"
		Goreauiella auriculata	"	"	"	"	Acanthostrongyle*	60 - 68	2.3 - 2.7	"
	Tab	Tabulospongia wellsi	Hartman & Goreau 1975	Stacked lamellar	Mg-cc	E.	Tylostyle* Spiraster**	c.290 Highly variable	c.3.5	" "
		Tabulospongia horiguchii	Mori 1976	"	"	"	Fusiform oxea* Sphaeraster form**	300 - 350	5.7 - I4.0 20 - 25	" "
		Tabulospongia japonica	Mori 1977	"	"	"	Dichotriaene*	300 - 355	I40 - I90	"
	Unas	Merlia normani	Kirkpatrick I909	Fascic.fib.	A	"	Tylostyle* Clavidisc** Raphide**	c.I40 c. 45 c. 80	c.I.8 c. 30 slender	" " "
		Merlia sp.	Hartman & Goreau 1970	"	"	"	Tylostyle* Clavidisc**	I60 35	I.8 26	" "
		Astrosclera willeyana	Kirkpatrick I9I0	Agg.spher.	"	"	Acanthostyle*	c. 70	c. 8	"
MESOZOIC L.Cretaceous	Mur	Murania lefeldi	Kazmierczak I974	Microgran. & fib.	?	I.m.	Style or Acanthostyle*	200	30	cc.
	Ch	Chaetetopsis favrei	Kazmierczak I979	Microgran.	?	I.s.	"	c.400	c.28	pyr.
	Cer	Neuropora pustulosa	Kazmierczak & Hillmer 1974	Fibro-normal	?A	"	?Acanthostyle*	I28 - I4I	6.6 - 7.6	cc.
U.Jur	Unas	Ptychochaetetes sp.	Termier & Termier 1976	"Lepidoporoid" (Scaly,porous)	?	I.d.	Monaxon*	?90	?I0	Fe ox.
U. Triassic	Cer	Keriocoelia conica	Dieci et al.1977	Fascic.fib.	?	I.m.	Style*	I90 ± 40	5.2 ± I.8	cc.
		Meandripetra zardinii	"	"	?	"	"	390 ± I40	27 ± 7	pyr.
		Sclerocoelia hispida	"	"	?	"	Acanthostyle*	6I ± I6	2.3 ± 0.7	cc.
		Sclerocoelia fasciculata	"	"	?	"	"	43 ± I2	2.3 ± 0.7	cc.
	?Cer	Atrochaetetes medius	"	"	?	"	"	74 ± 20	3.9 ± 0.9	cc.
PALAEOZOIC U.Carb	?Stp	Parallelopora mira	Newell 1935	Granular	?	I.s.	?Monaxon* (?Microscleres transverse to megascleres)	c. 250	c.I0	"
L.Carb	Ch	Chaetetes (Boswellia) mortoni sp.nov.	This paper	Fascic.fib.	?A	I.m.	Tylostyle* ?Raphide**	275 ± 50 c. 70	6.9 ± 0.9 c.3	Silica;cc.; pyr.
Dev	Stp	Stromatopora centrotum	Twitchell 1929	?	?	I.r.	"Spinose rod"*	c.I00	c.7	cc.

surface (Pl. 102, fig. 6). These colonies are calcitic, with a variable degree of microstructure alteration and compaction distortion. Spicule pseudomorphs have been observed in five of these colonies.

Comparison with material in the British Museum (Natural History) (repository prefix BMNH), Royal Scottish Museum (repository prefix R.S.M.) and Merseyside County Museum (repository prefix LIV.C.M.), led to the discovery of a further five spicule-bearing specimens of the same species. Two of these are partly silicified with intramural spicules locally replaced by silica.

SYSTEMATIC PALAEONTOLOGY

Class DEMOSPONGIAE Sollas, 1875
Sub-Class SCLEROSPONGIDEA Hartman and Goreau, 1972
Order CHAETETIDA Okulitch, 1936
Family CHAETETIDAE Milne-Edwards and Haime, 1850
Subfamily CHAETETINAE Milne-Edwards and Haime, 1850
Genus *CHAETETES* Fischer von Waldheim, 1830
Subgenus *BOSWELLIA* Sokolov, 1939

Type species. Chaetetes (*Boswellia*) *boswelli* Heritsch, 1932, 'Upper *Dibunophyllum* Zone (D_2)' of Ivovik, Serbia, U.S.S.R.

Diagnosis. Chaetetids with thickened irregular walls and rounded corners to lumina that may be either irregular or subpolygonal. Increase by pseudoseptal and basal fission. Incomplete fission and separation of pseudosepta into isolated columns occurs locally. Fascicular fibrous walls. Complete tabulae, variable in distribution. Intramural spicules (originally siliceous) present in some.

Remarks. Palaeozoic chaetetids have been subdivided generically on gross calicle morphology (Sokolov 1939, 1962). Sokolov (1939, p. 411) erected the subgenus *Chaetetes* (*Boswellia*) to include chaetetids with 'thickened irregular' calicle walls and 'undulate rounded' lumina, also stating that *C.* (*Boswellia*) 'occupies an intermediate position between . . .' *Chaetetes* and the meandrine genus *Chaetetipora* Struve, 1898. Species of *C.* (*Boswellia*) show this intermediate relationship very clearly, from the more prismatic thick-walled *C.* (*B.*) *uniformis*, Spiro 1961, and *C.* (*B.*) *heritschi* Sokolov, 1950, to the irregular calicles of *C.* (*B.*) *torquis* Spiro, 1961, which is very similar to some of the less meandrine chaetetiporinids, e.g. *Chaetetipora agonia* Sokolov, 1950. Although this classification is accepted provisionally here the division of the Chaetetidae at a generic level requires further clarification, that must now be based on an understanding of poriferan growth and variation.

EXPLANATION OF PLATE 102

Chaetetes (*Boswellia*) *mortoni* sp. nov., Lower Asbian (Lower Carboniferous, Eglwyseg Escarpment, Llangollen (Clwyd, North Wales).

Fig. 1. Paratype BMNH R49965. Compaction-fractured colony with laminar overgrowth. Negative of longitudinal section, ×2·3.

Fig. 2. Detail of a compaction-fractured thin-wall growth zone, BMNH R44965. Longitudinal section, ×45.

Fig. 3. Holotype BMNH R49964. Transverse section illustrating the subpolygonal to slightly irregular calicle pattern, ×12.

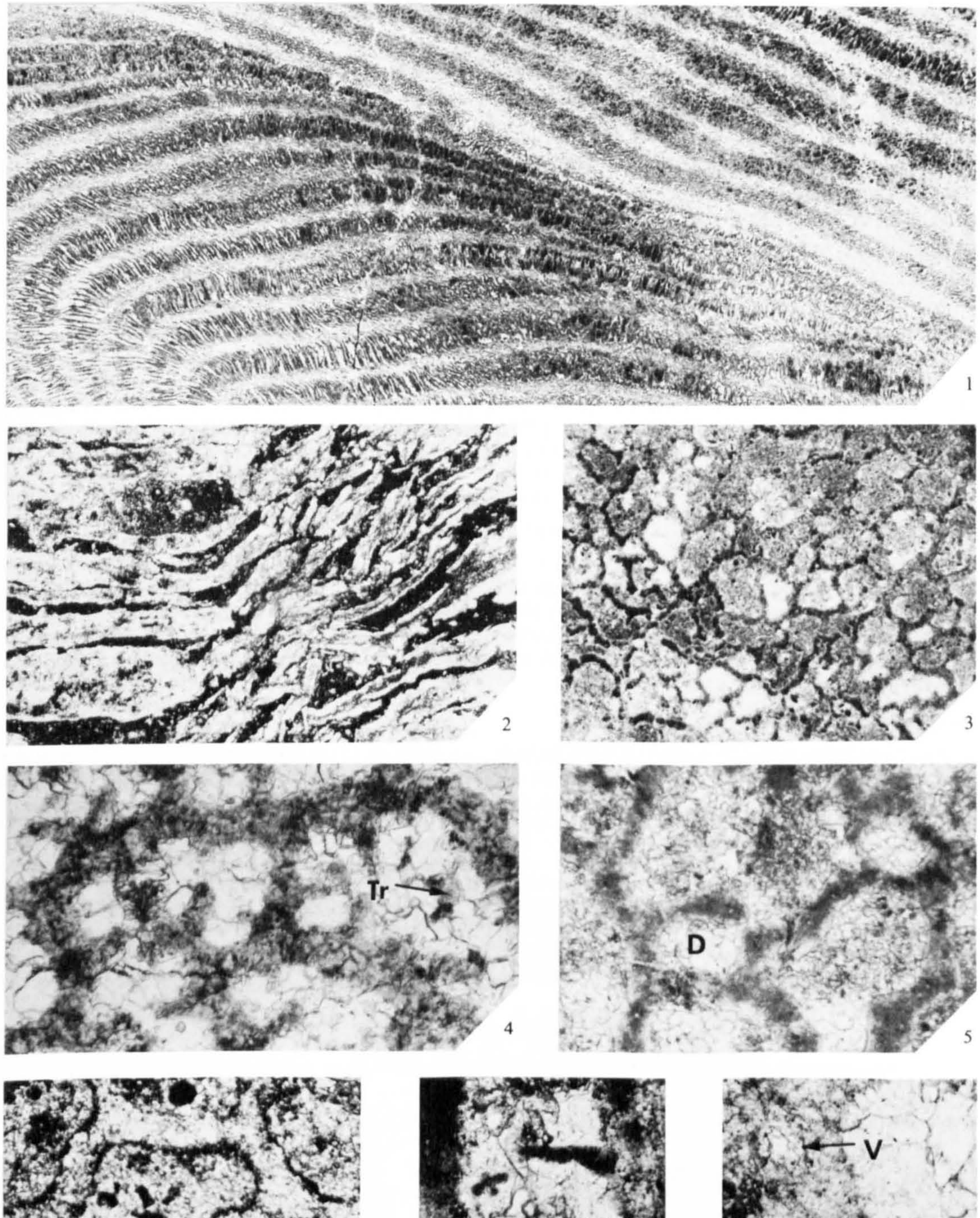
Fig. 4. Transverse section illustrating an irregular calicle pattern, with many pseudosepta, scalloped margins to lumina, and isolated trabecular column (Tr), BMNH R49965, ×35.

Fig. 5. Transverse section illustrating pseudoseptal fission in the corner of a calicle, BMNH R49964, ×40 (D = Daughter calicle).

Fig. 6. Transverse section illustrating mosaic neomorphic fabric to calicle walls with a micritic rim. Outer margin of BMNH R49964, ×70.

Fig. 7. Fractured tabula extending into lumen, longitudinal section, BMNH R49964, ×150.

Fig. 8. Sub-spherical 'vacuole' (V) within the calicle wall of BMNH R50133, ×100.



GRAY, Palaeozoic chaetetid

Chaetetes (*Boswellia*) *mortoni* sp. nov.

Plates 102, 103; text figs. 1–4; Table 2

Derivation of species name. After G. H. Morton who devoted many years of research to the Carboniferous of north Wales in the latter part of the nineteenth century.

Holotype. BMNH R49964, Tynant Limestone (Lower Asbian), quarried face, 400 m north of Tynant Ravine, 4 km north of Llangollen, Clwyd (National Grid Ref. SJ 21964573)

Paratypes. BMNH R4429 (Morton Collection), Lower Brown Limestone (in part equivalent to Tynant Limestone), Llangollen, Clwyd (partly silicified); BMNH R49965, Tynant Limestone (Lower Asbian), quarried face 500 m north of Tynant Ravine (SJ 21974582).

Other material. BMNH R50134, Tynant Limestone (Lower Asbian), World's End, 6 km north of Llangollen, Clwyd (SJ 23314789); BMNH R50133, BMNH R50135, and BMNH R50136 loose on scree slopes near Tynant Ravine, near the base of the Eglwyseg escarpment; BMNH R50188 and BMNH R50189, Tynant Limestone, 300 m north of Llwyn Hên-parc Gulley, Eglwyseg escarpment (SJ 22152638); LIV.C.M. 1974. 57, Eglwyseg Escarpment, Llangollen (in scree, partly silicified); BMNH R45851, Lower Carboniferous, Ravenstonedale, Cumbria; BMNH R46144, J. S. Baker Collection, Carboniferous Limestone (Blue Quarries), Ashfell Edge, Ravenstonedale, Cumbria; R.S.M. 1967.66.86–89 Nicholson Collection (thin sections only; all probably from one colony), Carboniferous Limestone, Archer Beck, Dumfriesshire.

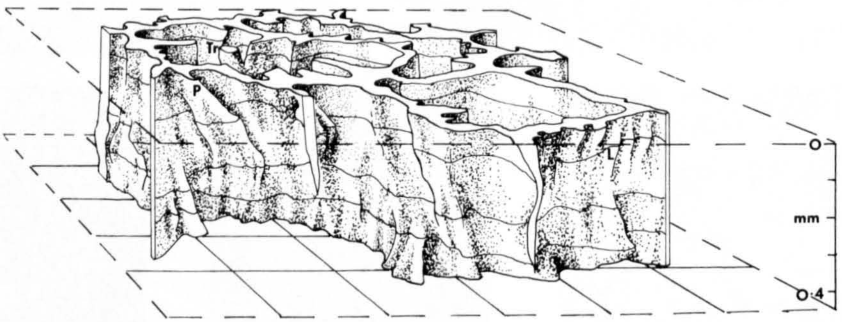
Range. ?Holkerian (BMNH R46144 from ?Ashfell Limestone) to Lower Asbian.

Diagnosis. Chaetetes (*Boswellia*) with irregular to subpolygonal (intracolonially variable) calicles. Fascicular fibrous walls with pseudosepta and irregular longitudinal ridges. Isolated pseudoseptal columns locally. Lumen diameter av. *c.* 500 μm; wall thickness av. *c.* 110 μm. Intramural spicules (monaxon tylostyle megascleres) subparallel the fibres, diverging distally, with their pointed (oxeote) ends directed distally. Spicule diameter *c.* 7 μm; spicule length variable *c.* 275 μm in well-preserved specimens. Tabulae well spaced. Basal and pseudoseptal fission only.

Description

Colony form. The colonies are laminar or bulbous, rarely greater than 12 cm diameter by 8 cm high. They display distinct growth bands in rhythms 2 to 5 mm thick, with zones of thinner calicle walls preferentially compaction fractured (Pl. 102, figs. 1, 2). On weathered and polished surfaces the thick-wall bands stand prominent, being a paler shade of brown-grey than the compaction fractured zones. Neither epitheca, astrorhizae, nor surface mamelons have been observed on any colony.

Calicle morphology. The calicles are irregular to subpolygonal (Pl. 102, figs. 3, 4), with the mean diameter of more polygonal lumina between 420 μm and 535 μm (see Table 2). The walls vary greatly in mean thickness, from *c.* 90 μm to *c.* 140 μm (measurements taken between ridges and pseudosepta). Pseudosepta are common, occasionally separating from the calicle walls as isolated columns. Both pseudosepta and ridges (undeveloped pseudosepta) longitudinally ornament the calicle walls (text-fig. 1) imparting a scalloped appearance to the lumina in



TEXT-FIG. 1. Block diagram of *Chaetetes* (*Boswellia*) *mortoni* sp. nov. to illustrate the development of longitudinal ridges (L), pseudosepta (P), and isolated trabecular columns (Tr) off the calicle walls. Diagram constructed from serial acetate peels of BMNH R50133.

transverse section (Pl. 102, fig. 4). Increase is by both pseudoseptal and basal fission. Pseudoseptal fission commonly occurs in calicle corners (Pl. 102, fig. 5). Incomplete pseudoseptal fission locally forms an irregular calicle pattern. Tabulae are rarely visible, appearing well spaced ($\leqslant$ 2 per mm), although this may in part be due to the degree of compaction fracture (Pl. 102, fig. 7).

Microstructure. The walls are fascicular fibrous penicillate calcite or chalcedonic silica, with a brown 'dusty' appearance in thin section due to submicroscopic to micrometre-sized inclusions of ?organic material. In the calcitic specimens these inclusions vaguely define the wall fibres and cause a variable pseudopleochroism (between paler and darker brown) cf. Hudson (1962). One specimen, BMNH R50133 has rare subspherical 'vacuoles', *c*. 50-μm diameter, within the calicle walls, of uncertain origin (Pl. 102, fig. 8). Neomorphism has destroyed details of the microstructure to varying degrees (Pl. 102, fig. 6; text-fig. 5), resulting in inclusion-poor areas lacking spicule relicts (especially the thin-wall growth bands). The walls rarely show a coarser fibrous fabric, with each fibre surrounded by thin brown pellicles that are probably the remnants of the ?organic inclusions.

Spicule form. In the calcitic specimens, spicule pseudomorphs occur within the walls subparallel to the fascicular fibres, diverging distally as straight or slightly curved elongate rods of clearer, inclusion-deficient calcite (Pl. 103, fig. 3) up to 300 μm long. In transverse section they appear as clear calcite circles or ellipses with a range of mean diameters between 6·6 μm and 8·2 μm (Pl. 103, fig. 4). Rarely they may exceed 20 μm diameter. Although surface detail is not visible on these pseudomorphs, their clarity varies from prominent to indistinct, reflecting variation in neomorphism. Rarely the spicules may be preserved as pyrite pseudomorphs with aggregates of pyrite crystals along their length (text-fig. 4*c*), similar to those described by Kaźmierczak (1979). In contrast some calicles of BMNH R4429 and LIV.C.M. 1974.57 are partially replaced by chalcedonic silica (Pl. 103, figs. 1, 2), with perfect intramural silica spicule pseudomorphs occurring adjacent to more vague calcitic ones. These pseudomorphs are low-relief colourless to high-relief red-brown translucent tylostyles, with circular cross-sections (Pl. 103, fig. 6), distinct bosses at their proximal ends and distally tapering points (Pl. 103, figs. 2, 8, 9, 10). As with the totally calcitic specimens these spicules diverge distally in the calicle walls (Pl. 103, fig. 2), subparalleling the fascicular fibres, but occasionally cross-cutting the wall-fibre trend at a high angle. Fossilized early corrosion features are seen on many spicule pseudomorphs (Pl. 103, figs. 7, 8, 10, 11).

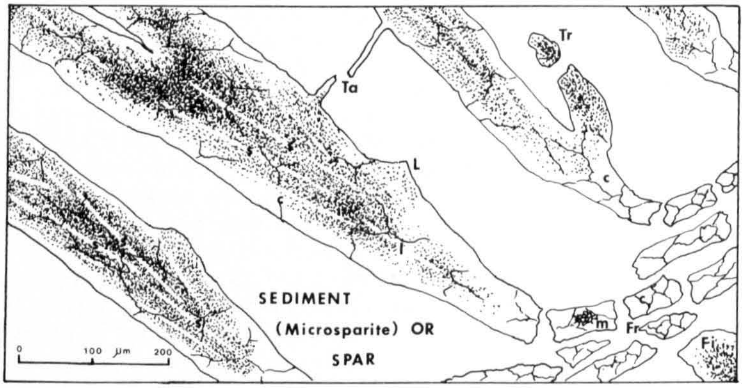
Seldom, and within BMNH R4429 only, ?raphide microsclere pseudomorphs occur (Pl. 103, fig. 7), as thin rods, pointed at both ends, *c*. 70 μm by 3 μm. These are only discernible in the chalcedonic regions, and would be too small to distinguish within the microstructure of calcitic calicle walls.

Spicule distribution. Spicule pseudomorphs have a variable distribution within the colonies. In the calcitic specimens they are only discernible within some of the ?organic-inclusion-rich areas associated with the less-fractured thick-wall growth bands where neomorphism has most perfectly replaced the primary fabric (see text-fig. 2). Specimens BMNH R4429 and LIV.C.M. 1974.57 also have a variable spicule distribution (see text-fig. 3) dependent on the replacement mineralogy. In BMNH R4429 totally and partly calcitic calicles rarely have visible calcitic spicule pseudomorphs. Here the microstructure is masked by a dense inclusion distribution. Siliceous spicules occur where the outer zone of these calicle walls is silicified. Only two growth bands are completely silicified (text-fig. 3), in which the best examples of a dense spicule distribution are visible (Pl. 103, figs. 1, 2).

TABLE 2. Variation in calicle and spicule size in six specimens of *Chaetetes* (*Boswellia*) *mortoni*

	Average lumen diameter in μm			Average wall thickness in μm			Average spicule length in μm			Average spicule width in μm		
SPECIMEN	mean	s.d.	n.	mean	s.d.	n.	mean	s.d.	n.	mean	s.d.	n.
*BMNH R49964	455	130	15	115	35	15	170	70	12	6.6	1.0	20
**BMNH R4429	420	140	10	94	29	10	275	50	10	6.9	0.9	10
**BMNH R49965	485	160	25	130	40	50	155	45	15	7.5	4.0	50
BMNH R50133	520	150	15	117	23	12	154	65	10	7.2	1.8	15
BMNH R50134	423	140	10	141	34	10	207	60	10	8.2	1.1	10
BMNH R50135	535	110	10	116	28	15	(Strongly neomorphosed microstructure)					

* = Holotype; ** = Paratype; s.d. = standard deviation; n. = sample number.



TEXT-FIG. 2. Sketch showing the relationships of the microstructural fabrics in calcitic specimens of *Chaetetes* (*Boswellia*) *mortoni* sp. nov. in longitudinal section. Symbols: Ta = fractured tabula; Tr = isolated trabecular column; L = longitudinal ridge or pseudoseptum; s = calcitic spicule pseudomorph; Fr = fracture zone of thin-wall growth bands; I = transition from ?organic-inclusion-dense thick-wall bands to inclusion-poor compaction-fractured zone; c = neomorphic crystal mosaic, with varying degrees of undulose and sweeping extinction often not discernible in the ?organic-inclusion poor regions; m = localized neomorphic fabric of vaguely fibrous microspar-size crystals, with thin brown pelicles.

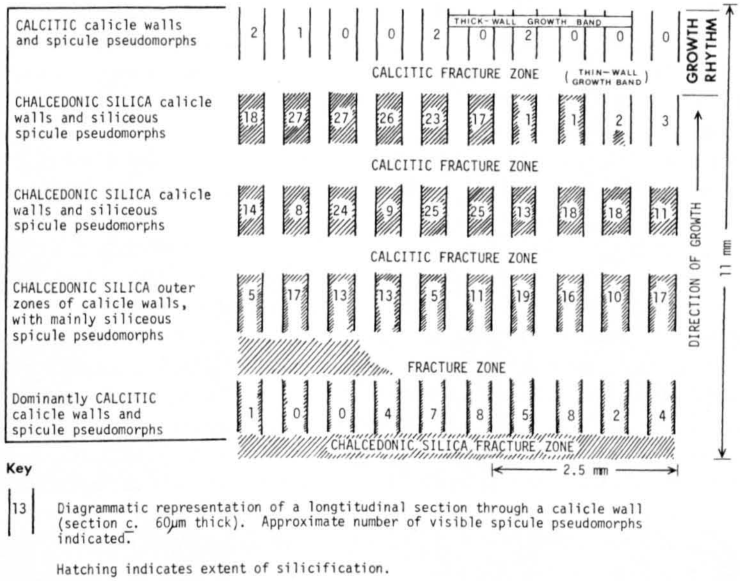
Discussion

The over-all similarity in colony, calicle, and microstructure form confirm that the calcitic and siliceous specimens are conspecific (see Table 2).

Species of *Chaetetes* are poorly defined. The size and degree of variation in calicle form, wall thickness, and tabula density are used as species-dependent characters. Many or all of these characters may, however, be controlled by environmental influences (e.g. Weyer 1967), and therefore where possible care must be taken to sample as large a population as possible. Variation within the specimens of *C.* (*B.*) *mortoni* described here is mostly intracolonial. Some similarities exist between this and previously described species. The calicle morphology is locally (Pl. 102, fig. 3) similar to *C.* (*B.*) *uniformis* Spiro, 1961 from the 'Visean' of the Moscow region, but differs by having a higher density of pseudosepta and slightly smaller calicles. Other closely comparable species are *C.* (*B.*) *torquis* Spiro, 1961, a densely tabulate form that displays many ridge swellings on its calicle walls (Spiro 1961, pl. 3, fig. 1*a*) in a similar manner to *C.* (*B.*) *mortoni*, *Chaetetipora agonia* Sokolov, 1950 and the larger-calicled *C. dubjanskyi* Sokolov, 1950. The latter two both show pseudoseptal fission in calicle corners (Sokolov 1950, pl. 15), but differ from *Chatetes* (*B.*) *mortoni* with their indistinctly meandrine form and dense tabula distribution. Compaction-fracture of *C.* (*B.*) *mortoni* colonies, however, imparts a superficial meandrine calicle pattern which may obscure the true calicle shape.

Chaetetipora etheridgeii (Thomson, 1881) is a variably meandrine species, characterized by a variable calicle shape, some arranged in 'sub-stellate groups radiating irregularly around a large central . . .' calicle (Thomson 1881, p. 208). It is densely tabulate, with calicles commonly from 0·5 to 2·0 mm diameter, and with thin calicle walls. *Chatetes* (*B.*) *mortoni* may be readily distinguished from this chaetetid by its lack of sub-stellate calicle-groups and its rare tabulae.

Whether the presence of intramural spicule pseudomorphs is a species-dependent factor is open to question. Extant sclerosponges have growth-variable spicule distributions (Hartman and Goreau 1972, p. 213) and Stearn (1972), remarked on a whole population of the extant *Astrosclera* Lister, 1900 (unassigned sclerosponge), from the Pacific without spicules. The problem is further complicated by the variable preservation of the spicules as pseudomorphs. In three calcitic specimens



TEXT-FIG. 3. Schematic diagram illustrating the spicule distribution within the calicle walls of the partly silicified BMNH R4429. Note that the higher numbers of visible spicule pseudomorphs per section of calicle wall occur in the chalcedonic silica zones, which are themselves in part controlled by the growth and fracture banding within the colony.

of *C.* (*B.*) *mortoni* they are undetected (BMNH R50135, BMNH R50136, and BMNH R50189). Therefore, although the spicular character is of great significance in understanding the phylogeny and histology of chaetetids, it must only be used with caution as a specific character in fossil forms.

MINERALOGY AND DIAGENESIS OF C. (*B.*) *MORTONI*

Basal calcareous skeleton. Extant sclerosponges secrete both aragonite (e.g. ceratoporellids) and high-magnesian calcite (e.g. tabulosponges) in their basal skeletons. Both are possible original mineralogies for Palaeozoic chaetetids. Fossil ceratoporellids (Viezer and Wendt 1976), probable sclerosponges (Dieci, Russo, and Russo 1974*a*), and stromatoporoids (Wendt 1975), that have retained their original aragonitic mineralogy, and have suffered little diagenetic alteration (Scherer 1977), have all been recorded from the Upper Triassic.

The *in situ* transformation of aragonite to calcite (Bathurst 1964; Dodd 1966), observed in Pleistocene scleractinian corals and molluscs (James 1974; Pingitore 1976; Wardlaw, Oldershaw, and Stout 1978), produces a secondary fabric which retains some detail of the primary microstructure. This transformation apparently occurs via a thin solution film less than 15 nm wide (Wardlaw *et al.* 1978, p. 1864) or a chalky solution zone (James 1974; Pingitore 1976). In this polymorphic transformation, relict detail of microstructure is defined by organic, and rarely aragonite, inclusions enclosed within a coarse mosaic of brown neomorphic calcite. Each of these coarse mosaic crystals exhibits straight (James 1974, p. 793) or undulose (Schneidermann, Sandberg, and Wunder, 1972, p. 88) extinction under crossed polars.

The transformation of high to low-magnesian calcite involves a paramorphic incongruent dissolution process (Plummer and Mackenzie 1974, p. 79). Fine detail is preserved in skeletal components during this transformation (e.g. Towe and Hemleben 1976), although ultrastructural changes may be noted (e.g. Sandberg 1975). In comparison, Lohmann and Meyers (1977, p. 1086) described milky skeletal calcite rich in microdolomite inclusions as evidence of an original high-magnesian calcite mineralogy, apparently caused by an 'incongruent dissolution or solid-stage exsolution' process re-equilibrating the magnesium within coarse crystals that acted as closed or semi-closed systems (Meyers and Lohmann 1978) during the mineralogical transformation. Richter and Fuchtbauer (1978) used the preservation of primary structures by ferroan calcite as a criterion for recognition of primary, high-magnesian calcite.

In calcitic specimens of *C.* (*B.*) *mortoni* a relationship between the crystal form and ?organic-inclusion distribution is observed. The paler-brown calicle walls of specimens with less included material comprise 50 μm to 300 μm mosaic crystals with irregular margins, and either straight or slightly undulose extinction under crossed-polars. Most inclusion-rich areas have sweeping or undulose extinction and some lack a crystal mosaic. Often the wall crystals are continuous with the clear lumen-filling spar. No distinct wall-fibre boundaries are visible but vague fibre boundaries are defined by trains of inclusions. Similarly, the margins of calcitic spicule pseudomorphs are indistinct and often masked by these inclusions. Compaction fractured zones lack a dense inclusion distribution and have a neomorphic mosaic which suggests that the transformation to calcite locally destroyed much of the original microstructure. Larger surface areas of fractured calicle walls, exposed to the calcifying pore waters, may have induced a more rapid and destructive mineralogical transformation.

In partly silicified specimens, the chalcedonic silica cementation and replacement occurred after compaction fracture, as indicated by the preferential silicification of their fracture zones. According to Meyers (1977), such compaction fracture could occur as a result of overburden pressure after burial to metres or tens of metres.

The ?organic-inclusions within the calicle walls appear to have provided a template retaining some detail of the original microstructure. The inclusion distribution may be partly related to the diagenetic history of the calicle walls as they are invariably less dense in compaction-fractured zones, and partly primary, caused by a variable secretion of organic matrix within the basal calcareous skeleton. This reliance on inclusions to define the primary microstructure, and the presence of neomorphic crystals with irregular margins, that often continue into the lumina as clear spar suggests an *in situ* transformation from primary aragonite to calcite. Schneidermann *et al.* (1972, p. 89) stated that continuity of neomorphic crystal fabrics from skeletal components into surrounding spar indicated an early aragonite void-cementation that could be 'expected to appear only in association

---

EXPLANATION OF PLATE 103

Chaetetes (*Boswellia*) *mortoni* sp. nov., Lower Asbian (Lower Carboniferous), Eglwyseg Escarpment, Llangollen (Clwyd, North Wales).

Fig. 1. Paratype BMNH R4429 (Morton Collection). Longitudinal section of calicle walls replaced by chalcedonic (length fast) silica, and zoned dolomite lumen infills. Microgranular silica spicule pseudomorphs visible as dark streaks within the calicle walls, × 50.

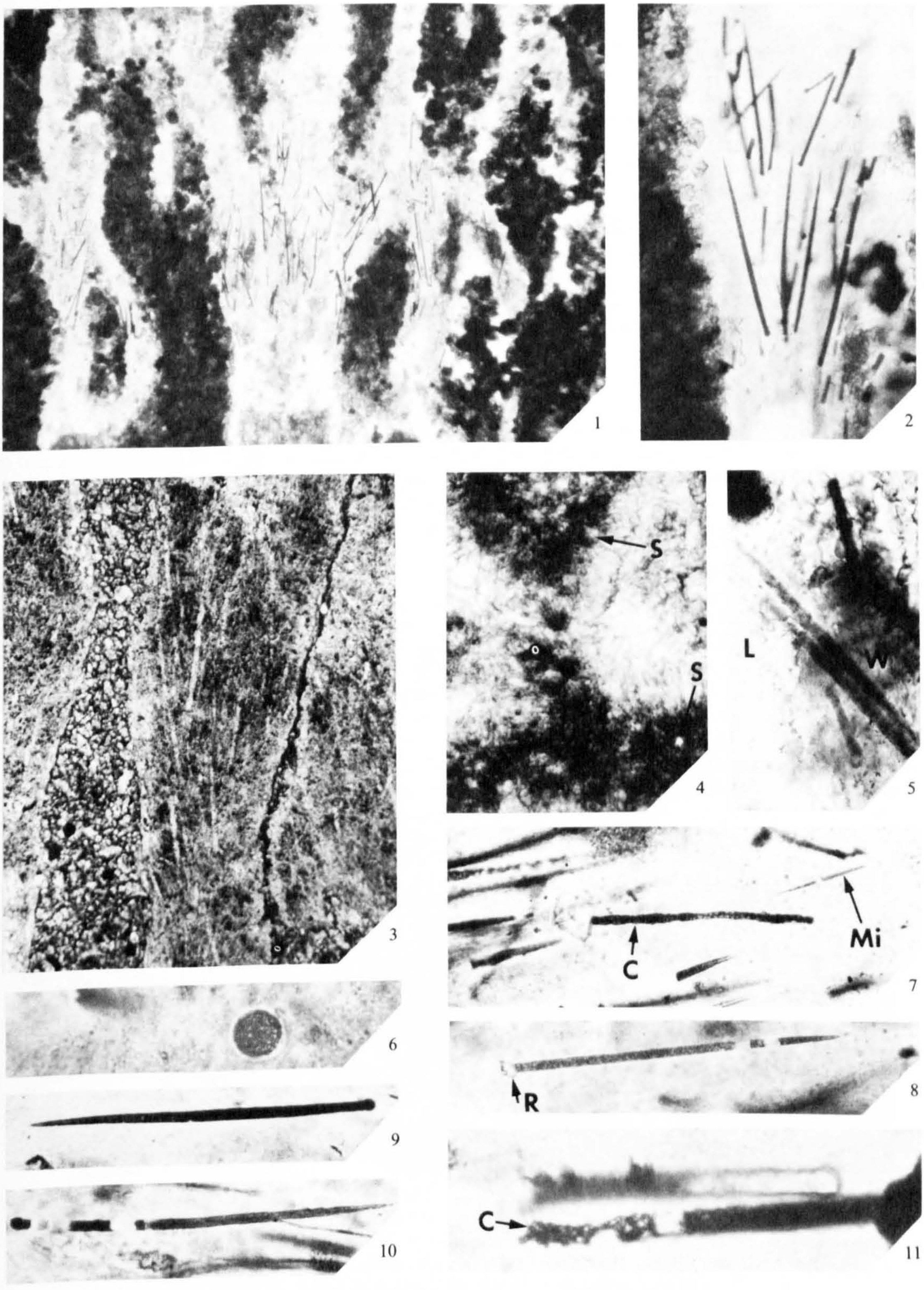
Fig. 2. BMNH R4429. Detail of chalcedonic silica calicle wall in longitudinal section, showing the spicule pseudomorphs diverging distally, × 150.

Fig. 3. Holotype BMNH R49964. Longitudinal section of calicle wall illustrating well-preserved calcitic spicule pseudomorphs, diverging distally and subparalleling the fascicular fibres, defined by trains of inclusions, × 120.

Fig. 4. BMNH R50134. Transverse section of calicle, with distinct and vague transverse sections through calcitic spicule pseudomorphs (S), × 150.

Fig. 5. BMNH R4429. Microgranular silica spicule pseudomorph extending to lumen void (L) calicle wall (W) junction, indicating original extension of distal portion of spicule into the lumen, and its subsequent dissolution, × 200.

Figs. 6–11. Microgranular silica spicule pseudomorphs of BMNH R4429. 6, transverse section, × 1000. 7, slight surface corrosion on tylostyle (C), with adjacent ?raphide-microsclere (Mi), × 200. 8, tylostyle with discontinuity and replacement by dolomite rhomb (R), × 200. 9, perfect tylostyle pseudomorph. Note proximal boss and distal point, × 200. 10, discontinuous tylostyle pseudomorph (probably a dissolution feature), × 200. 11, detail of a highly corroded tylostyle pseudomorph (C), × 600.



GRAY, Palaeozoic chaetetid

with previously aragonitic skeletons'. Conversely, in silicified specimens rare microdolomite inclusions occur (Pl. 103, fig. 8) within the calicle walls suggesting a high-magnesian calcite original mineralogy. However, intense dolomitization of the lumina of these colonies, the lack of microdolomite inclusions within calcitic specimens, and the common association of dolomite with chert nodules in the Tynant Limestone indicates that the magnesium may have an external source. Staining has not revealed any obviously 'ferroan' calcitic specimens. It would therefore appear that aragonite is the most probable original mineralogy of *C.* (*B.*) *mortoni*.

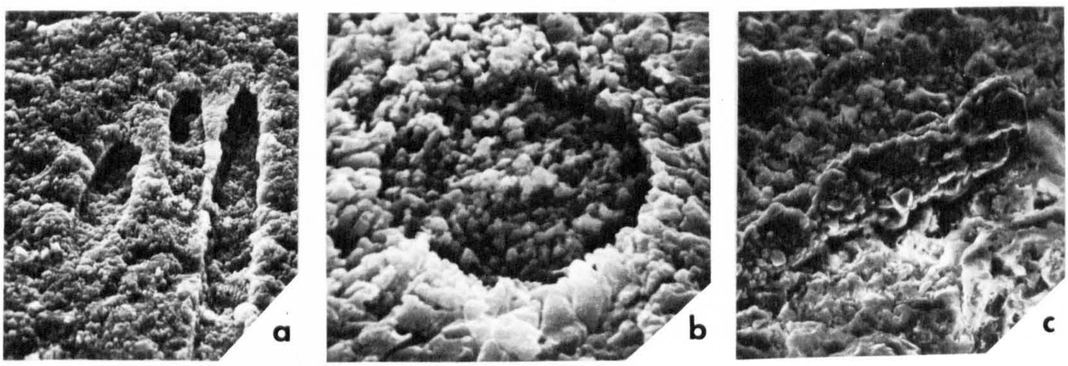
Spicule preservation. There are significant variations in the mode of preservation of the intramural spicules. Extant sclerosponges secrete siliceous spicules of various types (see Table 1) within or external to a calcareous skeleton. Hartman and Goreau (1970, pp. 210, 213) and Land (1976) described live colonies of ceratoporellans in which the opaline silica spicules (Opal 'A' of Jones and Segnit (1971)) were dissolving within the basal calcareous mass.

In *C.* (*B.*) *mortoni* spicules are preserved as calcite, silica, and pyrite pseudomorphs. Most specimens contain calcitic pseudomorphs. No spicule pseudomorphs convincingly extend into the lumina, although many spicules would have originally done so, as shown by extrapolation of spicule microstructure where it terminates abruptly against the calicle wall edge (Pl. 103, fig. 5).

The siliceous spicule pseudomorphs of BMNH R4429 and LIV.C.M. 1974.57 show dissolution features (Pl. 103, figs. 7, 11) from slight surface pitting to deep corrosion. Often the spicule pseudomorphs are discontinuous (Pl. 103, figs. 8, 10). The origin of this last feature is uncertain, but may be a severe localized corrosion effect. Dissolved parts of spicules have been replaced by clear chalcedonic silica indicating that a degree of dissolution occurred prior to the silicification of calicle walls. A later diagenetic event is indicated by dolomite rhombs replacing both spicule pseudomorphs (Pl. 103, fig. 8) and chalcedonic-silica walls.

In specimen BMNH R4429 neither axial canals nor axial filaments are visible, even with scanning electron microscopy of HF etched specimens (text-fig. 4*b*) (cf. Schwab and Shore 1971), indicating that an internal alteration of the spicule mineralogy has occurred. This is further confirmed by the presence of sub-microscopic microgranules which impart a red-brown hue and high relief to the spicule pseudomorphs. Deeper coloration corresponds to a more dense microgranule distribution. They give the spicule surface a smooth but frosted appearance in transmitted light. S.E.M. with E.D.A.X. shows that the microgranules are siliceous, and indistinguishable from the surrounding chalcedonic-silica walls. HF etched surfaces reveal the microgranule's form (text-fig. 4*b*). They vary between 0·2 μm and 0·5 μm diameter, and have sharp edges implying an internal structural ordering.

At an early or intermediate stage of diagenesis, biogenic opal 'A' is either converted *in situ* to opal



TEXT-FIG. 4. S.E.M. photomicrographs of spicule pseudomorphs in *C.* (*B.*) *mortoni*: 4*a*, longitudinal section of calicle wall with siliceous spicule pseudomorphs preferentially etched (10% HF for 3 minutes), BMNH R4429, $\times$ 1000; 4*b*, transverse section of a microgranular silica spicule pseudomorph (etched in 10% HF for 3 minutes), BMNH R4429, $\times$ 5000; 4*c*, pyritic spicule pseudomorph, composed of pyrite crystal aggregates (HCl etch), BMNH R46144, $\times$ 1000.

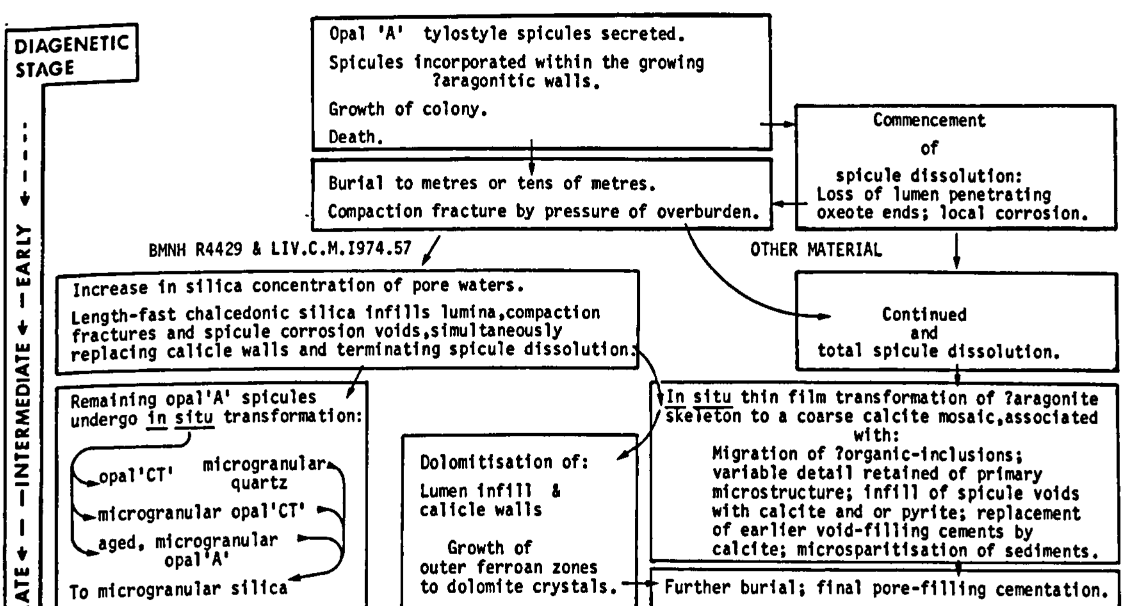
'CT' (Ernst and Calvert 1969; Wise and Weaver 1974, p. 305), a disordered cristobalite-tridymite silica polymorph (Jones and Segnit 1971), or dissolved and subsequently reprecipitated as opal 'CT' or quartz, filling voids and replacing carbonate grains (von Rad, Riech, and Rösch 1978). With increasing diagenetic maturity (depth of burial; time; temperature controls) opal 'CT' is eventually converted to quartz (Reich 1979). There are few records of opal 'CT' from pre-Cretaceous sediments.

The mineralogical composition of the spicule pseudomorphs in *C.* (*B.*) *mortoni* invites discussion. Are the spicules not pseudomorphs but diagenetically aged opal 'A' relic spicules? Reich (1979, pp. 754-755, pl. 1, fig. 4) reported Eocene opal 'A' sponge spicules with a pronounced microgranular texture similar to that observed on HF etched spicules in BMNH R4429.

Opal 'CT' normally occurs as bladed microspherules or lepispheres, 3 μm to 15 μm diameter. Siliceous sponge spicules have been recorded replaced by lepispheres that subsequently have been inverted to quartz (Reich 1979, p. 742) leaving visible relics of the precursor lepisphere. Von Rad *et al.* (1978, p. 903, pl. 3, figs. 2, 3) figured lepispheres replaced by microgranular quartz, also with a similar texture to that observed in spicule pseudomorphs of BMNH R4429 (text-fig. 4*b*). No lepisphere relics are visible in these spicules. Robertson (1978, p. 25) suggested that 'domains' of opal 'CT' would 'presumably appear' within opal 'A' 'which had escaped early diagenetic dissolution, becoming increasingly numerous as solid state ordering proceeds'. These 'domains' may be analogous to the microgranules within the spicule pseudomorphs (i.e. opal 'CT' without a lepisphere stage) or they may be a microgranular quartz replacement of opal 'CT' microgranules.

Opal 'CT' is generally recognized to have a higher relief than opal 'A', but exceptions are known. The high relief of microgranule-dense spicules in BMNH R4429 and LIV.C.M. 1974.57 may be caused by internal reflections on the surfaces of the microgranules. The spicules dissolve in HF far more readily than the surrounding chalcedonic silica (text. fig. 4*a*). This may be due to their fine granular nature, but Robertson (1978, p. 22) found that opal 'CT' dissolved preferentially in HF with respect to chalcedonic silica.

Thus the origin of the microgranular silica within the spicule pseudomorphs is unclear. It may be diagenetically aged opal 'A'; microgranular quartz replacement of opal 'CT' or relic microgranular opal 'CT' that has not gone through a lepisphere stage (see text-fig. 5). The nature of the silica matrix



TEXT-FIG. 5. Chart illustrating the sequence of events in the diagenetic history of *Chaetetes* (*Boswellia*) *mortoni* sp. nov. in relation to the preservation of microstructures and spicule pseudomorphs.

to the microgranules is unknown. In spicule pseudomorphs with few microgranules the matrix is optically similar to the surrounding chalcedonic silica. The occurrence of this microgranular silica fabric within the spicule pseudomorphs and not within the surrounding chalcedonic silica walls, and the retention of delicate spicule corrosion features suggest that the spicules were not originally calcareous, but probably opal 'A', as in extant sclerosponges.

In specimens BMNH R46144, BMNH R50188, and BMNH R45851 pyritic spicule pseudomorphs occur, similar to those described by Kaźmierczak (1979). The pyrite replacement is imperfect, with microcrystalline pyrite (text-fig. 4*c*) forming discontinuous chains or aggregates of 1 μm to 7 μm crystals along the pseudomorph length. The non-pyritic parts of the pseudomorphs are replaced by calcite. In BMNH R45851 the spicules are pseudomorphed by pyrite only towards the outer edges of calicle walls, where they merge with highly pyritic lumen-filling sediment. In contrast, BMNH R50188 has void-filling spar in the lumina, with pyritic spicule pseudomorphs terminating abruptly at the calicle wall margins, indicating that pyritization occurred after at least partial spicule dissolution. Rickard (1970) suggested that framboidal pyrite may replace organic globules or infill gaseous vacuoles. The calcitic spicule pseudomorphs often contain ?organic-inclusions, which may play a role in the pyrite formation.

AFFINITIES OF *C.* (*B.*) *MORTONI* WITH CERATOPORELLIDS

Ceratoporellids secrete a basal calcareous skeleton of fascicular fibrous aragonite calicles. These are subsequently infilled with fibrous aragonite by the upward-growing basal pinacoderm (Hartman and Goreau 1970, 1972). The calicles of *Ceratoporella nicholsoni* are regular, rounded to polygonal, rarely with a meandrine form (Hartman and Goreau 1972, fig. 17) similar to that of *Stromatospongia* Hartman, 1969, another extant ceratoporellid genus. Unlike chaetetids, ceratoporellids do not secrete tabulae. However, as Hartman and Goreau (1972, p. 142) point out, 'the difference is one of degree', with the growth of tabulae representing periodic rather than continuous carbonate secretion from the basal pinacoderm of the sponge animal. *Atrochaetetes*, a Mesozoic ceratoporellid (see above), lacks a continuous calicle infill (Cuif and Fischer 1974) and exhibits intramural spicule pseudomorphs in at least one species (Dieci, Russo, and Russo 1977). Its backfill may have formed by periodic secretion, or by periodic distally directed movement of the living tissues, and may be a character intermediate between solid backfills and tabulae.

The calicle surfaces of ceratoporellids are often ornamented with arborescent processes, rounded knobs, and spines of aragonite. No detailed calicle surface is available on *Chaetetes* (*B.*) *mortoni* for comparison; however, longitudinal sections show the distal edges of the calicles as rounded, although pre-burial erosion may have enhanced this. Isolated aragonitic trabeculae grow within the soft tissue of *Ceratoporella*, and are subsequently incorporated within the calcareous walls (Hartman and Goreau 1972, p. 135). These may be compared to the trabecular columns within *C.* (*B.*) *mortoni* which remain isolated during growth. Surface mamelons and astrorhizae, evident in some specimens of ceratoporellids as a result of differential growth-rates beneath excurrent canal systems (Stearn 1975), are not present on studied specimens of *C.* (*B.*) *mortoni*.

Ceratoporellids increase by pseudoseptal division, and Palaeozoic chaetetids by both pseudoseptal and basal fission. Mesozoic chaetetids in contrast also increase by intramural offset, as do tabulosponges.

Opal 'A' spicules are secreted from scleroblast cells within the living tissue of ceratoporellids and are incorporated within the skeleton as the colony grows. Hartman and Goreau (1972, p. 134) state that the spicules of *Ceratoporella nicholsoni* 'entrapped in the aragonite tend to follow the orientation of the calcareous crystalline units that surround them'. This is very like *C.* (*B.*) *mortoni*. The proximal (basal) spicule heads in living ceratoporellids are embedded within an organic matrix (Hartman and Goreau 1970). Although there is no direct evidence for organic fibres surrounding the head of spicules in *Chaetetes* (*B.*) *mortoni* the presence of ?organic-inclusions indicates an intimate relationship between the organic, calcareous, and siliceous components of the skeleton. Hartman and Goreau (1970, p. 213) also note that there are regions of the calcareous skeleton of *Ceratoporella* *nicholsoni* devoid of siliceous spicules. Although spicule preservation is variable throughout the colonies of *Chaetetes* (*B.*) *mortoni*, the local variation in spicule distribution may in part be primary. Ceratoporellids secrete monaxons of various forms, although they are neither known with tylostyles, nor with microscleres. The size of *C.* (*B.*) *mortoni* tylostyles does, however, fall within the size range of known ceratoporellid spicules.

The ecology of extant sclerosponges is fundamentally different from that of fossil chaetetids. Extant ceratoporellids are commonly found in a complex association with serpulid worms in submarine caves and at depth on fore-reef slopes (Hartman and Goreau 1970), whereas Palaeozoic chaetetids are common open-shelf dwellers, often associated with shallow-water carbonates.

There are significant similarities between *C.* (*B.*) *mortoni* and ceratoporellids in colony, calicle, microstructure form, and spicule character, indicating a close phylogenetic relationship. One notable difference is the presence of true tabulae in *C.* (*B.*) *mortoni* and the solid calcareous calicle infill characteristic of extant ceratoporellids.

AFFINITIES OF *C.* (*B.*) *MORTONI* WITH OTHER SCLEROSPONGES

Tabulospongids (Hartman and Goreau 1975; Mori 1976, 1977) secrete a calicular basal skeleton of high-magnesian calcite with a lamellar microstructure. The calicles are partitioned by horizontal tabulae, and spiny processes project into the lumina. The distribution of organic fibrils within the skeleton of tabulospongids is documented by Hartman and Goreau (1975, p. 167). These nanometre-sized fibrils act both as a matrix for the calcitic skeleton, and are present within the soft tissues. In *Tabulospongia horiguchii*, Mori (1976, pl. 3, fig. 3) shows that calicle wall centres are richest in organic matrix, resembling the distribution of probable organic matter now visible in calcitic specimens of *C.* (*B.*) *mortoni*. This may be a relict primary texture in the latter.

Tabulospongids secrete siliceous spicules that are not incorporated within their basal calcareous skeleton, but remain within the surface tissues. These spicules have a very low fossilization potential. The spicules are a variety of complex forms, with both megascleres and microscleres secreted by the same colony (see Table 1).

Although the calcareous skeleton is similar in design to many chaetetids, the marked differences in microstructure and spicule form and distribution readily distinguish tabulospongids from such Palaeozoic chaetetids as *C.* (*B.*) *mortoni*.

Merlia is an unassigned extant sclerosponge that secretes a prismatic tabular basal aragonitic skeleton, partitioned horizontally by incomplete tabulae. A variety of siliceous spicules are secreted within the living tissue, but not incorporated within the skeleton. Each prismatic calicle is formed by the outgrowth and interlocking of flanges, set at 120°, off stout fibre fascicles which form the calicle corners (see Stearn 1975). The architecture of Merlia therefore is subtly different from that of chaetetids. In contrast the types of spicules secreted by Merlia are similar to those of *C.* (*B.*) *mortoni* (see Table 1) (tylostyles and raphides) indicating some histologic similarities between these sponge animals.

Some Mesozoic stromatoporoids also have similarities with *C.* (*B.*) *mortoni*. Schnorf (1960) and Yabe and Sugiyama (1935) described Lower Cretaceous and Upper Jurassic forms with clear areas within the walls which may be sites of intramural spicules (Hartman and Goreau 1970), or part of the primary calcareous microstructure (Fenninger and Flajs 1974). The calcareous skeleton of these sclerosponges resembles chaetetids in as much as they also possess calicles with tabulae and hollow lumina. They are a diverse group, however, and show many characters atypical of Palaeozoic chaetetids.

In addition Table 1 lists the other known forms of spicule-bearing sclerosponges. These are not closely comparable with the present material but indicate the variety of microstructural and morphological patterns thus far encountered within the Sclerospongidea.

CONCLUSIONS

C. (*B.*) *mortoni* sp. nov. is a Palaeozoic (Upper Dinantian) chaetetid. The calicle morphology of this chaetetid is variable between slightly irregular (chaetetiporinid) and subpolygonal, having irregularly thick fascicular fibrous walls, ornamented with longitudinal ridges and pseudosepta. These factors place it within the subgenus *Boswellia* Sokolov, 1949, a typical Palaeozoic chaetetid.

Comparison of the basal calcareous skeleton with extant and fossil sclerosponges, the presence of a neomorphic mosaic, and the dependence on ?organic-inclusions to define the primary microstructure which is variably preserved, suggest that the original calcareous mineralogy was aragonite. Text-fig. 5 summarizes the approximate sequence of diagenetic events related to the preservation of microstructures in the skeletons of *C.* (*B.*) *mortoni.*

Spicule pseudomorphs occur within the fascicular fibrous walls as long thin tylostyles, with distally diverging oxeote ends that often would have penetrated the lumen of the sponge animal. They occur now as calcite, pyrite, or silica pseudomorphs, their mineralogy dependent on the diagenetic history of the basal skeleton.

In calcitic specimens, the spicule pseudomorphs are preserved as clear calcite rods in regions of the calicle walls with a dense ?organic-inclusion distribution. They are absent from compaction-fractured zones and are more common in thick-wall zones of growth rhythms. More rarely, the spicules are defined by trains or aggregates of pyrite crystals.

Chalcedonic silica locally replaces the calicle walls, enveloping spicule pseudomorphs that retain detail of their tylostyle form. Early dissolution features are fossilized within these spicules. Voids formed by spicule dissolution are infilled with clear chalcedonic silica. Remaining spicules have undergone alteration to microgranular silica of three possible forms, either diagenetically aged opal 'A', microgranular quartz replacement of opal 'CT', or relic microgranular opal 'CT' that has not gone through a lepisphere stage.

The presence of intramural spicule pseudomorphs within an otherwise typical member of the Palaeozoic Chaetetida further supports the sclerosponge affinities of at least some members of this group. Comparison of this chaetetid with other sclerosponges indicates that the spicular character and calcareous microstructure is very similar to that of the Ceratoporellida. The secretion of tabulae in chaetetids, rather than the backfill of ceratoporellids, remains the distinguishing microstructural feature. Of the two previously described intramural spicule pseudomorph bearing Mesozoic 'chaetetids' *Atrochaetetes* Cuif and Fischer, 1974, may be regarded as an aberrant member of the Ceratoporellida rather than a chaetetid *s.s.*, on account of its discontinuous backfill.

Acknowledgements. I would like to thank: Dr. C. T. Scrutton (University of Newcastle upon Tyne) and Dr. M. E. Tucker (University of Newcastle upon Tyne) for their advice and constructive criticism through all the stages of preparation of this paper; Dr. B. R. Rosen, Mr. R. F. Wise, Ms. J. G. Darrell (Department of Palaeontology, British Museum (Natural History)), Dr. C. D. Waterston (Royal Scottish Museum), and Mr. P. Phillips (Merseyside County Museum) who brought fossil material to my attention; and Mr. G. Staines (University of Newcastle upon Tyne) for his technical assistance with the electron microscopy. This research was carried out during the tenure of a N.E.R.C. research studentship.

REFERENCES

BATHURST, R. G. C. 1964. The replacement of aragonite by calcite in the molluscan shell wall. *In* IMBRIE, J. and NEWELL, N. (eds.), *Approaches to palaeoecology*. New York, J. Wiley & Sons, Inc., pp. 357-376.

CUIF, J.-P. and FISCHER, J.-C. 1974. Étude systématique sur les Chaetetida du Trias de Turquie. *Annls Paléont. Invertebrés*, **60**, 3-23, pls. I-IV.

DIECI, G., RUSSO, A. and RUSSO, F. 1974*a*. Nota preliminare sulla microstruttura di spugne aragonitiche del Trias medio-superiore. *Boll. Soc. paleont. ital.* **13**, 99-107, pls. 32-37.

—— —— —— 1974*b*. Revisione del genere *Leiospongia* d'Orbigny (Sclerospongia triassica). Ibid. 135-146, pls. 51-53.

—— —— —— and MARCHI, M. S. 1977. Occurrence of spicules in Triassic chaetetids and ceratoporellids. Ibid. **16**, 229-238, pls. 1-3.

DODD, J. R. 1966. Processes of conversion of aragonite to calcite with examples from the Cretaceous of Texas. *J. sedim. Petrol.* **36**, 733-741.

ERNST, W. G. and CALVERT, S. E. 1969. An experimental study of the recrystallization of porcelanite and its bearing on the origin of some bedded cherts. *Am. J. Sci.* **267A**, 114-133.

FENNINGER, A. and FLAJS, G. 1974. Zur mikrostruktur rezenter und fossiler Hydrozoa. *Biomineralization*, **7**, 69-99.

FISCHER, J.-C. 1970. Révision et essai de classification des Chaetetida (Cnidaria) post-Paléozoïques. *Annls Paléont. Invertebrés*, **56**, 151-220, pls. A-F.

FISCHER VON WALDHEIM, G. 1830. *Oryctographie du gouvernement de Moscou*, Moscow, 1st edn. 1830, 2nd edn. 1837.

HARTMAN, W. D. 1969. New genera and species of coralline sponges (Porifera) from Jamaica. *Postilla*, **137**, 1-39.

—— and GOREAU, T. F. 1970. Jamaican coralline sponges: their morphology, ecology and fossil relatives. *Symp. zool. Soc. Lond.* **25**, 205-243.

—— —— 1972. *Ceratoporella* (Porifera: Sclerospongiae) and the chaetetid 'corals'. *Trans. Conn. Acad. Arts Sci.* **44**, 133-148.

—— —— 1975. A pacific tabulate sponge, living representative of a new order of sclerosponges. *Postilla*, **167**, 1-14.

HERITSCH, F. 1932. *Chaetetes* und *Caninia* aus dem Karbon von Ivovik bei Krupanj in West-Serbien. *Bull. Serv. Géol. Yougosl.* **1/2**, 1-10, pls. I-II.

HUDSON, J. D. 1962. Pseudopleochroic calcite in recrystallized shell-limestones., *Geol. Mag.* **92**, 492-500, pl. XXI.

JAMES, N. P. 1974. Diagenesis of scleractinian corals in the subaerial vadose environment. *J. Paleont.* **48**, 785-799.

JONES, J. B. and SEGNIT, E. R. 1971. The nature of opal. I. Nomenclature and constituent phases. *J. geol. Soc. Aust.* **18**, 57-68.

KAŹMIERCZAK, J. 1974. Lower Cretaceous sclerosponge from the Slovakian Tatra Mountains. *Palaeontology*, **17**, 341-347, pls. 45-46.

—— 1979. Sclerosponge nature of chaetetids evidenced by spiculated *Chaetetopsis favrei* (DENINGER 1906) from the Barremian of Crimea. *Neues Jb. Geol. Paläont. Mh.* **1979**, 97-108.

—— and HILLMER, G. 1974. Sclerosponge nature of the Lower Hauterivian 'bryozoan' *Neuropora pustulosa* (ROEMER 1839) from western Germany. *Acta palaeont. Pol.* **19**, 443-453.

KIRKPATRICK, R. 1909. Notes on *Merlia normani* Kirkp. *Ann. Mag. nat. Hist.* **(8) 4**, 42-48.

—— 1910. On the affinities of *Astrosclera willeyana* Lister. Ibid. **(8) 5**, 380-383.

—— 1912*a*. *Merlia normani* and its relation to certain Palaeozoic fossils. *Nature, London.* **89**, 502-503.

—— 1912*b*. On the nature of stromatoporoids. Ibid. 607.

LAND, L. S. 1976. Early dissolution of sponge spicules from reef sediments, North Jamaica. *J. sedim. Petrol.* **46**, 967-969.

LÉVI, C. 1973. Systématique de la classe des Demospongia (Démosponges). *In* GRASSÉ, P. (ed.). *Traité de Zoologia*, Masson, Paris. Vol. 3, 577-3631.

LOHMANN, K. C., and MEYERS, W. J. 1977. Microdolomite inclusions in cloudy prismatic calcites: A proposed criterion for former high-magnesium calcites. *J. sedim. Petrol.* **47**, 1078-1088.

MEYERS, W. J. 1977. Chertification in the Mississippian Lake Valley Formation, Sacramento Mountains, New Mexico. *Sedimentology*, **24**, 75-105.

—— and LOHMANN, K. C. 1978. Microdolomite-rich syntaxial cements: proposed meteoric-marine mixing zone phreatic cements from Mississippian limestones, New Mexico. *J. sedim. Petrol.* **48**, 475-488.

MILNE-EDWARDS, H. A. and HAIME, J. 1850-1854. A monograph of the British fossil corals. *Palaeontogr. Soc. (Monogr.)* Parts **1-5**, i-lxxxv, 1-322, pls. I-LXII.

MORI, K. 1976. A new recent sclerosponge from Ngargol, Palau Islands and its fossil relatives. *Sci. Rep. Tôhoku Univ.*, 2nd ser. (Geol.), **46**, 1-9, pls. 1-6.

—— 1977. A Calcitic sclerosponge from the Ishigaki-shima coast, Ryukyu Islands, Japan. Ibid. **47**, 1-5, pls. 1-2.

MORTON, G. H. 1879. *The Carboniferous Limestone and Cefn-y-Fedw Sandstone of the country between Llanymynech and Minera North Wales*, 140 pp. David Bogue, London.

NEWELL, N. D. 1935. Some mid-Pennsylvanian invertebrates from Kansas and Oklahoma: II. Stromatoporoidea, Anthozoa and Gastropoda. *J. Paleont.* **9**, 341-355, pls. 33-36.

OKULITCH, V. J. 1936. On the genera *Heliolites*, *Tetradium* and *Chaetetes*. *Am. J. Sci.* **32**, 361-379.

PINGITORE, N. E. JR 1976. Vadose and phreatic diagenesis: Process products and their recognition in corals. *J. sedim. Petrol.* **46**, 978-984.

PLUMMER, L. N. and MACKENZIE, F. T. 1974. Predicting mineral solubility from rate data: application to the dissolution of magnesian calcites. *Am. J. Sci.* **274**, 61-83.

RAD, U. VON, RIECH, V. and RÖSCH, H. 1978. Silica diagenesis in continental margin sediments off Northwest Africa. *In* LANCELOT, Y., SEIBOLD, E., *et al.* 1978. *Initial reports of the deep sea drilling project.* Washington (U.S. Government Printing Office), **41**, 879-905.

RIECH, V. 1979. Diagenesis of silica, zeolites, and phyllosilicates at sites 397 and 398. *In* VON RAD, U., RYAN, W. B. F., *et al.* 1979. *Initial reports of the deep sea drilling project.* Washington (U.S. Government Printing office), **47**, 741–759.

RICHTER, D. K. and FUCHTBAUER, H. 1978. Ferroan calcite replacement indicates former magnesian calcite skeletons. *Sedimentology*, **25**, 843–860.

RICKARD, D. T. 1970. The origin of framboids. *Lithos*, **3**, 269–293.

ROBERTSON, A. H. F. 1977. The origin and diagenesis of cherts from Cyprus. *Sedimentology*, **24**, 11–30.

SANDBERG, P. A. 1975. Bryozoan diagenesis: bearing on the nature of the original skeleton of rugose corals. *J. Paleont.* **49**, 587–606.

SCHERER, M. 1977. Preservation, alteration and multiple cementation of aragonitic skeletons from the Cassian Beds (U. Triassic, Southern Alps): petrographic and geochemical evidence. *Neues Jb. Geol. Paläont. Abh.* **154**, 213–262.

SCHNEIDERMANN, N., SANDBERG, P. A., and WUNDER, S. J. 1972. Recognition of early cementation of aragonitic skeletal carbonates. *Nature phys. Sci., Lond.* **240**, 88–89.

SCHNORF, A. 1960. Parastromatoporoidae nouveaux du Jurassique supérieur et du Valanginien inférieur du Jura. *Eclog. geol. Helv.* **53**, 729–732, pls. I–III.

SCHWAB, D. W. and SHORE, R. E. 1971. Fine structure and composition of a siliceous sponge spicule. *Biol. Bull. mar. biol. Lab., Woods Hole*, **140**, 125–136.

SCRUTTON, C. T. 1979. Early fossil cnidarians. *In* HOUSE, M. R. (ed.), *Spec. Vol. Syst. Ass.* **12**, 161–207.

SOKOLOV, B. S. 1939. Stratigraphical value and types of Chaetetidae of the Carboniferous of the USSR. *C.r. (Dokl.) Acad. Sci. URSS* **23**, 409–412.

—— 1950. Khetetidy karbona severo-vostochnoi Ukrainy i sporedel'nykh oblastei (s opisaniem nekotorykh tabulyat). [Carboniferous Chaetetidae of the Northeast Ukraine and adjacent areas.] *Trudȳ vses. neft. nauchno-issled. geol.-razv. Inst.* N.S. **27**, 1–101, pls. I–XX. [In Russian.]

—— 1962. Gruppa Chaetetida. *In* ORLOV, Y. A. (ed.), *Osnovy Paleontologii*: Porifera, Archaeocyatha, Coelenterata, Vermes. Moscow [in Russian]. *Fundamentals of Paleontology*, **2**, 259–270. *Israel Prog. sci. Transl.*, Jerusalem 1971 [English translation].

SOMERVILLE, I. D. 1979. A sedimentary cyclicity in early Asbian (Lower D1) limestones in the Llangollen district of North Wales. *Proc. Yorks. geol. Soc.* **42**, 397–404.

SPIRO, T. A. 1961. Nekotorie Vizeiskie y Namyurskie chetetidi Podmoskovbya. [Some Viséan and Namurian chaetetids from the Moscow region.] *Paleont. Zh.* **4**, 24–28, pls. II–III. [In Russian.]

STEARN, C. W. 1972. The relationship of the stromatoporoids to the sclerosponges. *Lethaia*, **5**, 369–388.

—— 1975. The stromatoporoid animal. *Lethaia*, **8**, 89–100.

TERMIER, H. and TERMIER, G. 1976. Spongiaires hypercalcifiés et ectoproctes sténolèmes fossiles. *C.r. Acad. Sci. Paris*, **282D**, 1269–1272.

THOMSON, J. 1881. On the genus *Alveolites*, *Amplexus* and *Zaphrentis*, from the Carboniferous System of Scotland. *Proc. Phil. Soc. Glasgow*, **13**, 194–237.

TOWE, K. M. and HEMLEBEN, C. H. 1976. Diagenesis of magnesian-calcite: evidence from miliolacean foraminifera. *Geology*, **4**, 337–339.

TWITCHELL, G. B. 1929. The structure and relationships of the true stromatoporoids. *Am. Midl. Nat.* **11**, 270–306, pls. XVIII–XXVI.

VACELET, J. 1977. Éponges Pharétronides actuelles et Sclérosponges de Polynésie francaise, de Madagascar et de la Réunion. *Bull. Mus. natn. Hist nat.* 3e sér., **444**, Zool. 307, 345–368.

—— VASSEUR, P. and LÉVI, C. 1976. Spongiares de la pente externe des récifs coralliens de Tulear (Sud-Ouest de Madagascar). *Mém. Mus. natn. Hist. nat.* **A44**, 116 pp.

VEIZER, J. and WENDT, J. 1976. Mineralogy and chemical composition of Recent and fossil skeletons of calcareous sponges. *Neues Jb. Geol. Palaont. Mh.* **1976**, 558–573.

WARDLAW, N., OLDERSHAW, A. and STOUT, M. 1978. Transformation of aragonite to calcite in a marine gastropod. *Can. J. Earth Sci* **15**, 1861–1866.

WENDT, J. 1975. Aragonitische Stromatoporen aus der alpinen Obertrias. *Neues Jb. Geol. Palaont. Abh.* **150**, 111–125.

WEYER, VON D. 1967. Einige Chaetetidae (Coelenterata) aus dem Unterkarbon des Velberter Sattels (Rheinisches Schiefergebirge). *Geologie*, **16**, 1156–1163, pls. I–II.

WISE, S. W. and WEAVER, F. W. 1974. Chertification of oceanic sediments. *In* HSU, K. J. and JENKYNS, H. C. (eds.). *Pelagic sediments: on land and under the sea. Spec. Publs int. Ass. Sediment.* **1**, 301–306.

YABE, H. and SUGIYAMA, T. 1935. Jurassic stromatoporoids from Japan. *Sci. Rep. Tôhoku imp. Univ.*, ser. 2 (Geol.), **14**, 135–191, pls. XL–LXXI.

DAVID I. GRAY

Department of Geology
University of Newcastle Upon Tyne
Newcastle Upon Tyne, NE1 7RU

Typescript received 12 September 1979
Revised typescript received 21 January 1980

THE PALAEONTOLOGICAL ASSOCIATION

The Association was founded in 1957 to further the study of palaeontology. It holds meetings and demonstrations as well as publishing *Palaeontology* and *Special Papers in Palaeontology*. Membership is open to individuals and to institutions on payment of the appropriate annual subscription. Rates for 1980 are:

Institutional membership	£25 (U.S. $60)
Ordinary membership	£12 (U.S. $29)
Student membership	£7·50 (U.S. $18)

There is no admission fee. Correspondence concerned with Institutional Membership should be addressed to Dr. C. H. C. Brunton, Department of Palaeontology, British Museum (Natural History), Cromwell Road, London SW7 5BD, England. Student members are persons receiving full-time instruction at educational institutions recognized by the Council. On first applying for membership, an application form should be obtained from the Membership Treasurer. Subscriptions cover one calendar year and are due each January; they should be sent to the Membership Treasurer (for name and address see Council list below).

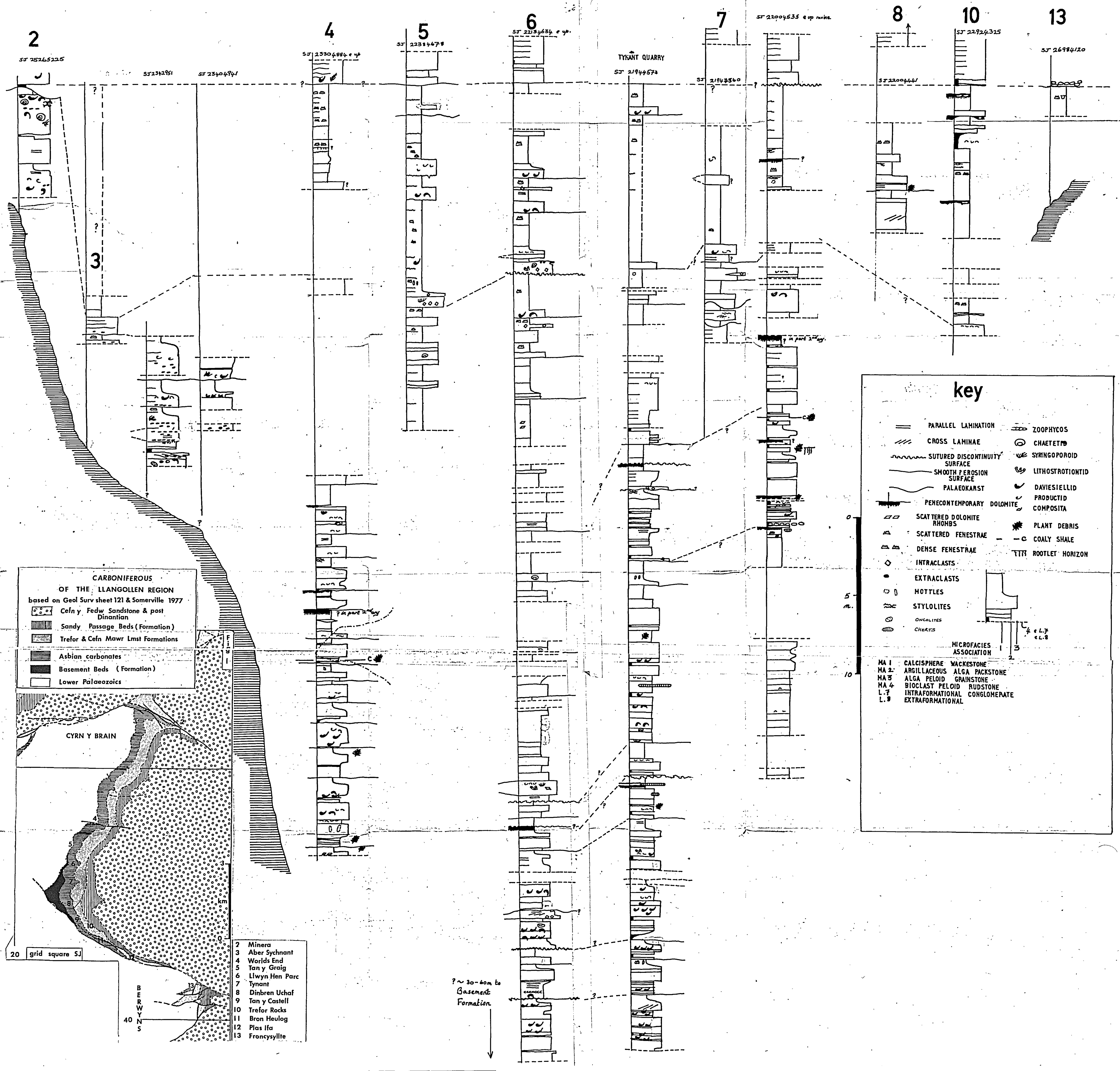
PALAEONTOLOGY

All members who join for 1980 will receive Volume 23, Parts 1–4. All back numbers are still in print and may be ordered from B. H. Blackwell, Broad Street, Oxford OX1 3BQ, England, at £15 (U.S. $36) per part (post free).

SPECIAL PAPERS IN PALAEONTOLOGY

Members may subscribe to this by writing to the Membership Treasurer (for name and address see Council list below); the subscription rate for 1980 is £20 (U.S. $48) for Institutional Members, and £10 (U.S. $24) for Ordinary and Student Members. A single copy of each *Special Paper* is available to Ordinary and Student Members *only*, for their personal use, at a discount of 25% below the prices shown in the list inside the front cover. Non-members may obtain copies, at the listed prices, from B. H. Blackwell, Broad Street, Oxford OX1 3BQ, England.

CHART A TYNANT FORMATION LLANGOLLEN



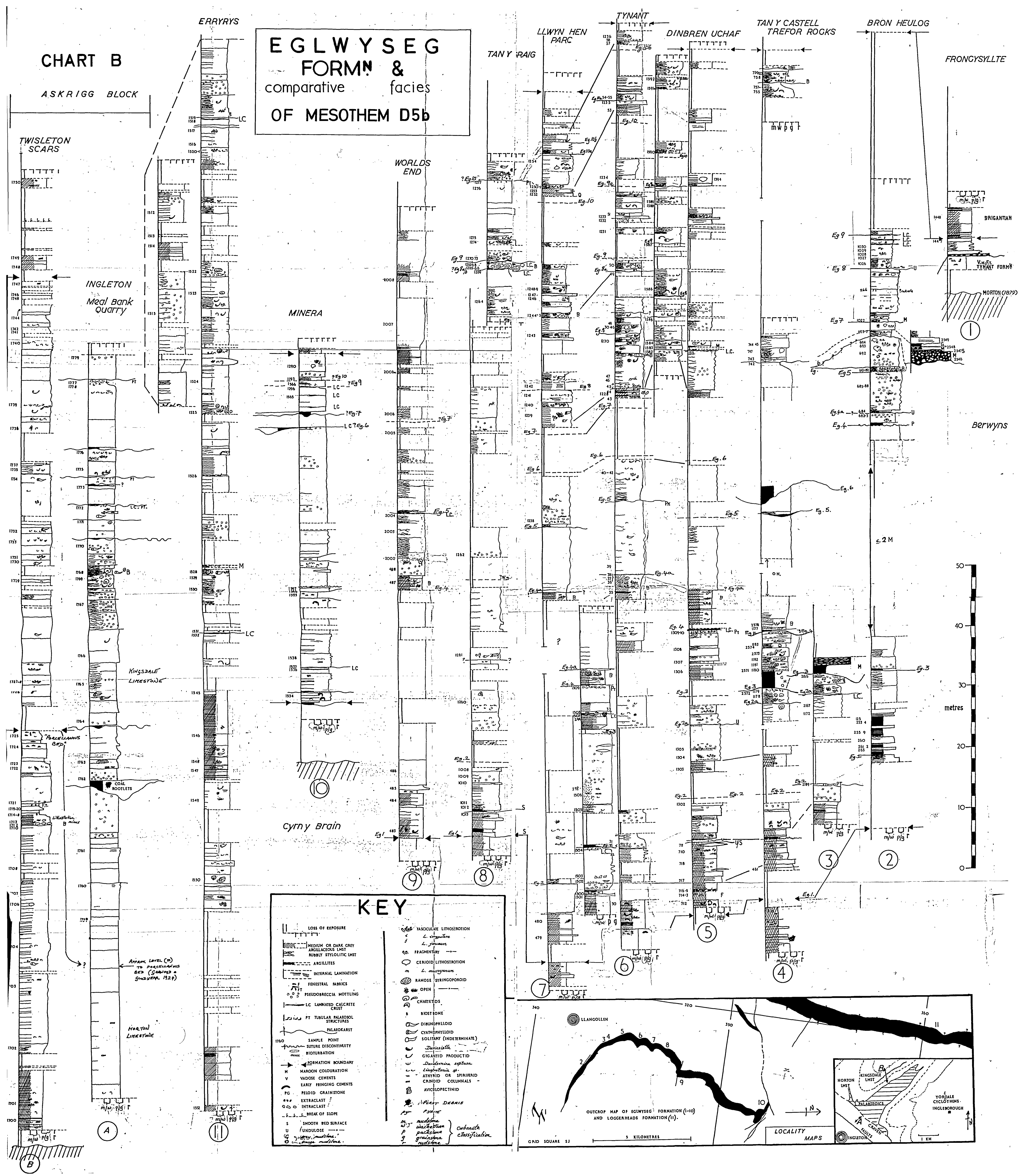


CHART B
EGLWYSEG FORMN. & comparative facies OF MESOTHEM D5b
ASKRIGG BLOCK
TWISLETON SCARS
INGLETON
Meal Bank Quarry
ERRYRYS
MINERA
WORLDS END
TAN Y RAIG
LLWYN HEN PARC
TYNANT
DINBREN UCHAF
TAN Y CASTELL
TREFOR ROCKS
BRON HEULOG
FRONCYSYLLTE
BRIGANTIAN
TYNANT FORMN
MORTON (1879)
Berwyns
KINGSDALE LIMESTONE
HORTON LIMESTONE
"PORCELLANOUS BED"
COAL ROOTLETS
APPROX LEVEL (?) TO PORCELLANOUS BED (GARWOOD & GOODYEAR 1924)
Cyrny Brain
metres
KEY
LOSS OF EXPOSURE
MEDIUM OR DARK GREY ARGILLACEOUS LMST
RUBBLY STYLOLITIC LMST
ARGILLITES
INTERNAL LAMINATION
FENESTRAL FABRICS
PSEUDOBRECCIA MOTTLING
LC LAMINATED CALCRETE CRUST
PT TUBULAR PALAEOSOL STRUCTURES
PALAEOKARST
SAMPLE POINT
SUTURE DISCONTINUITY
BIOTURBATION
FORMATION BOUNDARY
M MAROON COLOURATION
V VADOSE CEMENTS
EARLY FRINGING CEMENTS
PG PELOID GRAINSTONE
EXTRACLAST
INTRACLAST
BREAK OF SLOPE
S SMOOTH BED SURFACE
U UNDULOSE
FASCICULATE LITHOSTROTION
L. junceum
FRAGMENTARY
CERIOID LITHOSTROTION
L. mucosarium
RAMOSE SYRINGOPOROID
OPEN
CHAETETIDS
B BIOSTROME
DIBUNOPHYLLOID
CYATHOPHYLLOID
SOLITARY (INDETERMINATE)
Daviesiella
GIGANTID PRODUCTID
Davidsonina septosa
Linoproductus sp.
ATHYRID OR SPIRIFERID
CRINOID COLUMNALS
AVICULOPECTINID
PLANT DEBRIS
py PYRITE
m mudstone
w wackestone
p packstone
g grainstone
r rudstone
Carbonate Classification
OUTCROP MAP OF EGLWYSEG FORMATION (1-10) AND LOGGERHEADS FORMATION (11).
LLANGOLLEN
GRID SQUARE SJ
5 KILOMETRES
LOCALITY MAPS
HORTON LMST
KINGSDALE LMST
PALAEOZOICS
YOREDALE CYCLOTHEMS
INGLEBOROUGH
INGLETON
CRAVEN FAULTS
1 KM

CHART C
OSWESTRY ASBIAN

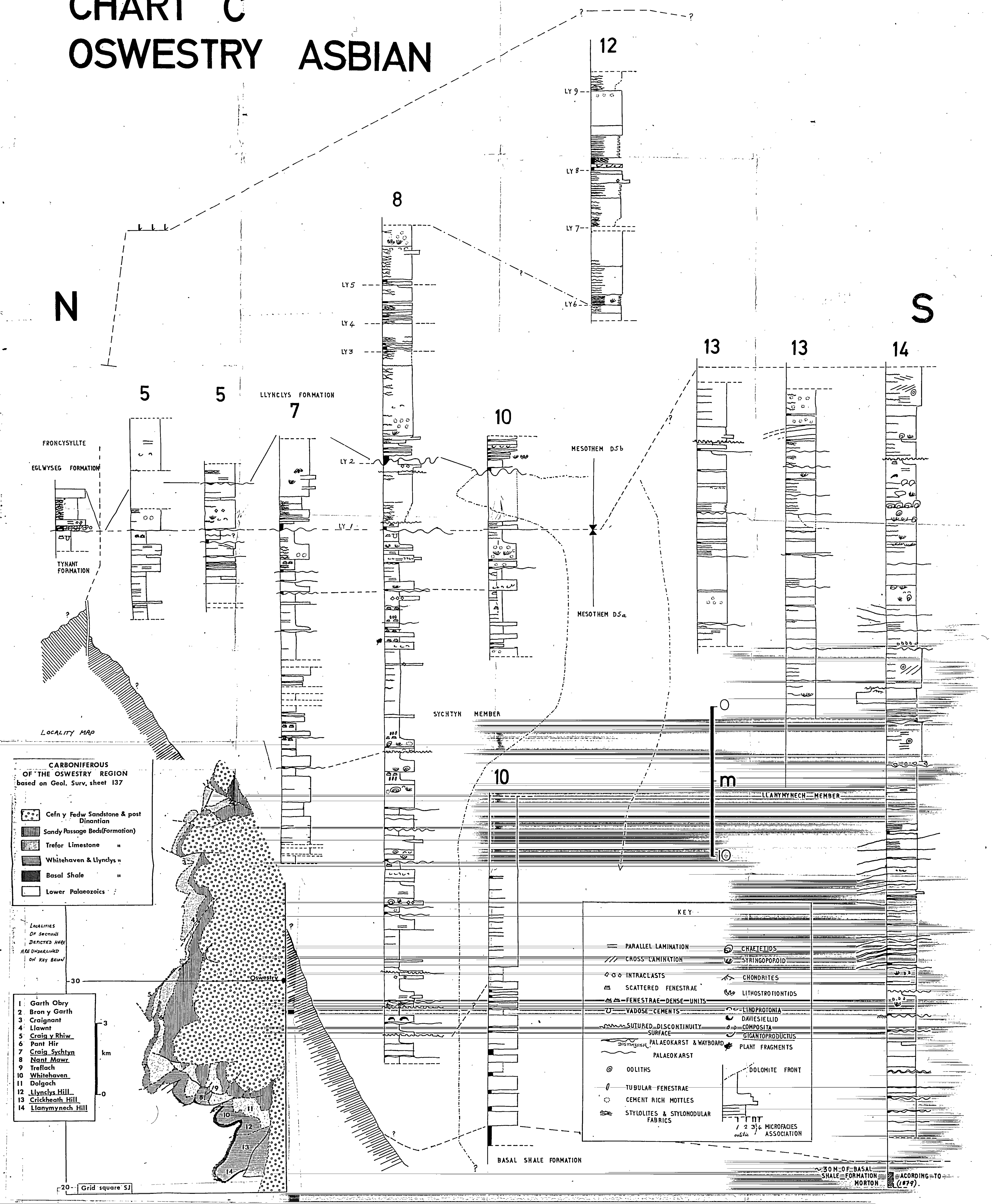


CHART D
TREFOR FORMATION
GRAIG quarry (CEFN MAWR LIMESTONE)
Lower part of succession
Biostrome
Base of Cefn Mawr Limestone
ASBIAN
Northern quarry exposures.
Quarry MINERA (southern quarry face)
EGLWYSEG FORMATION
Hafod MINERA
Succession approx 50 m here
SANDY PASSAGE FORMATION
20 m unexposed
259517
? Shale
CEFN Y CIST 237485
CYRN Y BRAIN
GRID SQUARE SJ
CARBONIFEROUS OF THE LLANGOLLEN REGION
based on Geol Surv sheet 121 & Somerville 1977
Cefn y Fedw Sandstone & post Dinantian
Sandy Passage Beds (Formation)
Trefor & Cefn Mawr Lmst Formations
Asbian carbonates
Basement Beds (Formation)
Lower Palaeozoics
BERWYNS
2 Minera
3 Aber Sychnant
4 Worlds End
5 Tan y Graig
6 Llwyn Hen Parc
7 Tynant
8 Dinbren Uchaf
9 Tan y Castell
10 Trefor Rocks
11 Bron Heulog
12 Plas Ifa
13 Froncysyllte
(MORTON'S CORAL BED)
EGLWYSEG PLANTATION
TYNANT
14 m unexp.
TREFOR ROCKS
? Shale
? Shale in part
PLAS IFA 246424
FRONCYSYLLTE
ca 10 m to U Asbian
GARTH-OBRY 258380
BRON Y GARTH 263373
SANDY PASSAGE FORMATION
CRAIGNANT
10-8 m unexposed
BRON-Y-GARTH SJ 249370
LAWNT
PANT HIR
1 Garth Obry
2 Bron y Garth
3 Craignant
4 Llawnt
5 Craig y Rhiw
6 Pant Hir
7 Craig Sychtyn
8 Nant Mawr
9 Treflach
10 Whitehaven
11 Dolgoch
12 Llynclys Hill
CARBONIFEROUS OF THE OSWESTRY REGION
based on Geol. Surv. sheet 137
Oswestry
Sandy Passage Formation
Trefor Limestone
Whitehaven & Llynclys
Basal Shale
Lower Palaeozoics
Dolgoch WHITEHAVEN
TREFLACH
Hafod-y-Calch CORWEN
Unexposed
Includes Orionastraea shale of Morton (1879) towards base
SANDY PASSAGE FORMATION
TREFOR FORMATION
metres
KEY
CRINOID FRAGS.
FASCICULATE COLONIES
Lithostrotion junceum
L. cf irregulare
L. pauciradiale
L. martini
Diphyphyllum spp
Nemistium edmondsi
CERIOID & PLOCOID COLONY
Lonsdaleia spp
Lithostrotion decipiens
SOLITARY CORAL
SYRINGOPOROID
CHAETETID
GIGANTID PRODUCTID
Semiplanus
PRODUCTID
Camarotoechia
Bio; B BIOSTROME
Girvanella
Saccaminopsis
FORAMINIFERA
UNGDARELLID
Girvanella nodule
Zoophycos
MINOR CYCLE BOUNDARIES ARE WITH PREFIX "Tf".
pk palaeokarst
sutured discontinuity surface
s SMOOTH SURFACE
u UNDULOSE SURFACE
LOADED HORIZON
SHALE
CALCAREOUS SHALE
L.C. LAMINATED CRUST
MOTTLE (CEMENT-RICH)
BURROW
GENERAL BIOTURBATION
PALE (ARGILLITE-POOR) SEDIMENT
DARK (ARGILLITE-RICH) SEDIMENTS
TUBULAR FENESTRAE
DOLOMITE & FRONT
pyr PYRITE
pel PELOIDAL
mar MARRON
QUARTZ CLASTS
pt TUBULAR CALCRETE STRUCTURES
umb. UMBER COLOURATION
sie. BURNT SIENNA COLOURATION
LAMINATION
Unc ANGULAR BEDDING DISCORDANCE

CHART E

VERTICAL AND LATERAL MICROFACIES VARIATIONS — MINERA QUARRY – ? STRATA BETWEEN Eg.1 & Eg.4