

**PALAEOENVIRONMENTAL EVOLUTION OF AN ORDOVICIAN-SILURIAN  
DEEP-MARINE SEDIMENTARY SUCCESSION IN THE WELSH BASIN**

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in the  
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by

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## ABSTRACT

The Llandeilo-Llandovery stratigraphic succession of west Wales comprises fourteen formations and consists of two dominant lithologies, namely sandstone and mudstone, the distribution of which is largely determined by eustatic sea-level changes. The sediments are interpreted as representing distal shelf, sandstone lobes, lobe fringes and slope environments. The turbidite systems developed within the Welsh depositional basin as a Type 1 (unchannelled sandstone lobes) system. The tectonic setting of the basin for the entire period of deposition was that of an active margin. Petrographic and geochemical data, however, indicate a passive margin tectonic setting for the basin. Critical examination of the models suggests that the sedimentary provenance indicators are incorrect and that the signature is a relict one, presumably derived from the Precambrian basement.

Laminated hemipelagites are recorded from a number of formations. These formed under anaerobic conditions within the basin. They all contain layered pyrite framboids which are interpreted as having formed diagenetically within the sediment. It is also possible that some of the framboids may have formed syngenetically in the water column.

The strata contain a diverse and relatively abundant ichnofaunal assemblage consisting of sixteen ichnogenera: Chondrites, Circulichnis, Cochlichnus, Cosmorhappe, Desmograption, Gordia, Helminthoida, Helminthopsis, Nereites, Neonereites, Palaeophycus, Paleodictyon (Glenodictyon), Paleodictyon (Squamodictyon), Planolites, Protopaleodictyon and Spirophycus. The ichnogenera are unevenly distributed throughout the succession, the main controlling factors being topography, anoxia and a global extinction event.

**TABLE OF CONTENTS**

**PAGE**

**CHAPTER ONE - INTRODUCTION**

1.1 Aims.....1.1  
1.2 Organisation.....1.2  
1.3 Methodology.....1.4  
1.4 References.....1.6

**CHAPTER TWO - SEDIMENTARY FACIES**

2.1 Introduction.....2.1  
2.2 Disorganised gravels, gravelly muds  
and pebbly sands.....2.1  
2.3 Organised gravels and pebbly sands.....2.2  
2.4 Sandstones.....2.3  
2.5 Sand-Mud couplets.....2.4  
2.6 Silts, muddy silts and silt-mud  
couplets.....2.5  
2.7 Muds and Clays.....2.6  
2.8 Chaotic deposits.....2.8  
2.9 References.....2.9

**CHAPTER THREE - STRATIGRAPHY AND SEDIMENTOLOGY OF THE DINAS  
HEAD TO NEW QUAY AREA**

Abstract.....3.1  
Part One:  
3.1 Introduction.....3.1  
3.2 Previous work.....3.2  
3.3 Stratigraphy.....3.3  
    3.3.1 Parrog Formation.....3.3  
    3.3.2 Newport Sands Formation.....3.5  
    3.3.3 Ceibwr Formation.....3.7  
    3.3.4 Poppit Sands Formation.....3.8  
Part Two:

3.4	Introduction.....	3.10
3.5	Gwbert Formation.....	3.10
3.6	Mwnt Formation.....	3.12
3.7	Tresaith Formation.....	3.14
3.8	Llangranog Formation.....	3.16
3.9	Gaerglwyd Formation.....	3.18
3.10	Allt Goch Formation.....	3.19
3.11	Cefn Cwrt Formation.....	3.20
3.12	Llangraig Formation.....	3.20
3.13	Grogal Formation.....	3.21
3.14	Aberystwyth Grits Formation.....	3.22
3.15	Summary.....	3.26
3.16	References.....	3.27

**CHAPTER FOUR - SEDIMENTARY ENVIRONMENTS OF THE  
LLANDEILO-LLANDOVERY COASTAL SUCCESSION**

Abstract.....	4.1
4.1 Tectonic setting and palaeogeography.....	4.1
4.1.1 Tectonic setting of the Welsh Basin.....	4.1
4.1.2 Palaeogeography of the Welsh Basin.....	4.4
4.2 Palaeoflow in the succession.....	4.7
4.2.1 Anomalous palaeocurrents to the north of Aberystwyth.....	4.8
4.3 Vertical sequence analysis.....	4.9
4.4 Facies Associations and their distributions..	4.10
4.4.1 Distal shelf facies.....	4.10
4.4.2 Slope facies.....	4.11
4.4.3 Sandstone lobe facies.....	4.12
4.4.4 Lobe fringe facies.....	4.14
4.5 Palaeogeography.....	4.15
4.5.1 Llandeilo-Caradoc palaeogeography.....	4.17
4.5.2 Caradoc-Ashgill palaeogeography.....	4.18
4.5.3 Early-Middle Llandovery palaeogeography.	4.19
4.5.4 Late Llandovery palaeogeography.....	4.20
4.6 Factors controlling turbidite system development.....	4.21
4.6.1 Eustatic changes in sea level.....	4.21
4.6.2 Tectonic setting.....	4.22

4.6.3	Sediment source.....	4.23
4.7	The Sea of Okhotsk as a modern analogue for the Welsh Basin.....	4.24
4.7.1	The Sea of Okhotsk.....	4.24
4.7.2	The Welsh Basin and the Sea of Okhotsk..	4.25
4.8	References.....	4.27

**CHAPTER FIVE - PETROLOGICAL AND GEOCHEMICAL DETERMINATION OF  
PROVENANCE IN THE SOUTHERN WELSH BASIN**

Abstract.....	5.1
5.1 Introduction.....	5.1
5.2 Tectonic history of the Welsh Basin.....	5.2
5.3 Age, previous work and depositional setting..	5.2
5.4 Sandstone petrography and framework modes....	5.4
5.5 Provenance and tectonic discrimination.....	5.5
5.6 Microconglomerate petrography.....	5.6
5.7 Chemical classification of rocks.....	5.7
5.8 Geochemical analysis of provenance.....	5.9
5.8.1 Quartz richness.....	5.10
5.8.2 Major element analysis of sandstones.....	5.10
5.8.3 K/Rb diagram.....	5.11
5.8.4 $K_2O/Na_2O$ vs $SiO_2$ diagram.....	5.11
5.8.5 Trace element concentrations.....	5.12
5.8.6 Multi-element diagram normalised to post-Archean average shale.....	5.13
5.9 Discussion.....	5.13
5.9.1 Sandstone provenance.....	5.15
5.9.2 Mudstone provenance.....	5.16
5.10 Conclusions.....	5.19
5.11 References.....	5.20

**CHAPTER SIX - LLANDEILO-LLANDOVERY LAMINATED HEMIPELAGITES  
FROM THE WELSH BASIN**

Abstract.....	6.1
6.1 Introduction.....	6.1

6.2	Location and stratigraphy.....	6.2
6.3	Laminated hemipelagites.....	6.3
6.4	X-ray diffraction analysis of Llandeilo- Llandovery sediments.....	6.4
6.5	Levels of organic carbon.....	6.5
6.6	Geochemistry of hemipelagites and fine- grained turbidites.....	6.5
6.7	Back-scattered electron microscopy of hemipelagites and associated fine-grained turbidites.....	6.6
6.8	Discussion.....	6.7
6.9	Conclusions.....	6.12
6.10	References.....	6.14

**CHAPTER SEVEN - DISTRIBUTION OF ORDOVICIAN-SILURIAN  
ICHTNOFOSSIL ASSEMBLAGES IN WALES - IMPLICATIONS  
FOR PHANEROZOIC ICHNOFAUNAS**

Abstract.....	7.1
7.1 Introduction.....	7.1
7.2 Location, lithologies and age of study area.....	7.2
7.3 Ichnofossil distribution in the Ordovician- Silurian succession of west Wales.....	7.3
7.3.1 Recurrent ichnofossil associations..	7.4
7.3.2 Palaeoenvironments and trace fossil associations.....	7.5
7.4 Factors affecting the ichnofossil distribution.....	7.5
7.4.1 Toponymy.....	7.5
7.4.2 Anoxia.....	7.6
7.4.3 Water depth.....	7.7
7.4.4 Global extinction events.....	7.9
7.5 Discussion and conclusions.....	7.11
7.6 References.....	7.16

**CHAPTER EIGHT - A NEREITES ASSEMBLAGE FROM THE  
ORDOVICIAN-SILURIAN SUCCESSION OF THE WELSH**

## DEPOSITIONAL BASIN

Abstract.....	8.1
8.1 Introduction.....	8.1
8.2 Location and stratigraphy.....	8.2
8.3 Depositional environment.....	8.3
8.4 Previous work.....	8.4
8.5 Systematic ichnology.....	8.5
8.6 Palaeoecology.....	8.44
8.6.1 Ichnofaunal distribution patterns within the Llandeilo-Llandovery succession.....	8.44
8.6.2 Ethology of the ichnofaunal assemblage..	8.48
8.7 Conclusions.....	8.48
8.8 References.....	8.50

## CHAPTER NINE - THE ICHNOGENUS DESMOGRAPTON FROM THE SILURIAN OF WALES - FIRST RECORD FROM THE PALAEOZOIC

Abstract.....	9.1
9.1 Introduction.....	9.1
9.2 Location and stratigraphy.....	9.2
9.3 Sedimentary environment.....	9.2
9.4 Systematic ichnology.....	9.3
9.5 References.....	9.7

## CHAPTER TEN - CONCLUSIONS

10.1 Conclusions.....	10.1
10.2 Suggestions for further work.....	10.3
10.3 References.....	10.5

## APPENDIX ONE - GEOCHEMICAL DATA FROM THE SOUTHERN WELSH BASIN

**LIST OF FIGURES**

**After page**

Fig. 1.1 Map of Wales showing study area.....1.2

Fig. 1.2 Map of study area showing previously  
published work.....1.2

Fig. 1.3 Regional and tectonic framework of the  
Welsh Basin.....1.3

Fig. 1.4 The "Variscan Front" and structures  
inferred to have been active during the  
early Palaeozoic history of the Welsh  
Basin.....1.4

Fig. 3.1 Stratigraphic map of the Dinas Head to New  
Quay area.....3.1

Fig. 3.2 Key to stratigraphic sections.....3.3

Fig. 3.3 Measured section of Parrog Formation,  
Parrog.....3.4

Fig. 3.4 Measured section of Newport Sands Formation,  
Newport Sands.....3.6

Fig. 3.5 Measured section of Ceibwr Formation,  
Ceibwr.....3.7

Fig. 3.6 Measured section of Poppit Sands Formation,  
Poppit Sands (Section 1a).....3.9

Fig. 3.7 Measured section of Poppit Sands Formation,  
Poppit Sands (Section 1b).....3.10

Fig. 3.8 Measured section of Poppit Sands Formation,  
Poppit Sands correlating sections  
1a and 1b.....3.10

Fig. 3.9 Measured section of Poppit Sands Formation,

	Rees Quarry.....	3.10
Fig. 3.10	Measured section of Gwbert Formation, Gwbert (Section 1).....	3.11
Fig. 3.11	Measured section of Gwbert Formation, Gwbert (Section 2).....	3.12
Fig. 3.12	Measured section of Mwnt Formation, Mwnt.....	3.13
Fig. 3.13	Measured section of Tresaith Formation, Aberporth (Section 1).....	3.15
Fig. 3.14	Measured section of Tresaith Formation, Aberporth (Section 2).....	3.15
Fig. 3.15	Measured section of Tresaith Formation, Tresaith (Section 3).....	3.16
Fig. 3.16	Measured section of Llangranog Formation, Llangranog.....	3.17
Fig. 3.17	Measured section of Gaeroglwyd Formation, Traeth-yr-Ynys.....	3.18
Fig. 3.18	Measured section of Allt Goch Formation, Llangranog.....	3.20
Fig. 3.19	Measured section of Grogal Formation, Cwm Tudu (Section 1).....	3.22
Fig. 3.20	Measured section of Grogal Formation, Craig Coybal (Section 2).....	3.22
Fig. 3.21	Measured section of Aberystwyth Grits Formation, New Quay (Section 1).....	3.23
Fig. 3.22	Measured section of Aberystwyth Grits Formation, New Quay Quarry (Section 2)....	3.24

Fig. 3.23	Measured section of Aberystwyth Grits Formation, New Quay Harbour (Section 3)...	3.25
Fig. 4.1	Regional and tectonic framework of the Welsh Basin.....	4.1
Fig. 4.2	The "Variscan Front" and structures inferred to have been active during the early Palaeozoic history of the Welsh Basin.....	4.3
Fig. 4.3	Palaeocurrent map of the study area.....	4.7
Fig. 4.4	Palaeocurrent directions in the Clarach to Borth region.....	4.8
Fig. 4.5	Block diagram showing a facies reconstruction for the study area during the Llandeilo and Caradoc.....	4.17
Fig. 4.6	Block diagram showing a facies reconstruction for the study area during the Caradoc and Ashgill.....	4.17
Fig. 4.7	Block diagram showing a facies reconstruction for the study area during the Early-Middle Llandovery.....	4.19
Fig. 4.8	Block diagram showing a facies reconstruction for the study area during the Middle-Late Llandovery.....	4.19
Fig. 4.9	Schematic stratigraphic column of the study area relating the lithological divisions to eustatic sea-level changes.....	4.21
Fig. 4.10	Topographic map of the Sea of Okhotsk and the Lower Palaeozoic Welsh Basin.....	4.24
Fig. 5.1	Location and simplified geological map of the study area.....	5.2

Fig. 5.2	Generalised stratigraphic column and sea-level curve for the study area.....	5.2
Fig. 5.3	Triangular QFL plot for sandstones from the southern Welsh Basin.....	5.5
Fig. 5.4	Triangular QmFLt plot for sandstones from the southern Welsh Basin.....	5.5
Fig. 5.5	Triangular QpLvMLsm plot for sandstones from the southern Welsh Basin.....	5.6
Fig. 5.6	Quartz-richness of southern Welsh Basin sandstones based on major element geochemistry.....	5.7
Fig. 5.7(a-c)	Harker variation diagrams of major elements for sandstones and mudstones from the southern Welsh Basin.....	5.7
Fig. 5.7(d-i)	Harker variation diagrams of major elements for sandstones and mudstones from the southern Welsh Basin.....	5.8
Fig. 5.8	Bivariate plots for the discrimination of plate tectonic setting of sandstones from the southern Welsh Basin.....	5.10
Fig. 5.9	Distribution of K (log wt%) and Rb (ppm) in the southern Welsh Basin sandstones relative to a K/Rb ratio of 230.....	5.11
Fig. 5.10	Tectonic discrimination diagram for mudstones from the southern Welsh Basin.....	5.11
Fig. 5.11	Multi-element diagram normalised to Post-Archean average shale.....	5.13
Fig. 5.12	Comparative diagram showing (a) present day tectonic setting of South America, and (b)	

Ashgill-Llandovery palaeogeographic reconstruction of the North Atlantic region.....5.15

- Fig. 6.1 Selected typical bulk-rock XRD traces for the sediments studied.....6.4
- Fig. 6.2 Bulk-rock XRD traces for silt-mud tripartite sequences from the Gwbert Formation.....6.5
- Fig. 6.3 Bulk-rock XRD traces for silt-mud tripartite sequences from the Tresaith Formation.....6.5
- Fig. 6.4 Bulk-rock XRD traces for mudstone turbidites and laminated hemipelagites from the Gaerlwyd Formation.....6.5
- Fig. 7.1 Generalised stratigraphic column and sea-level curve of the area.....7.3
- Fig. 7.2 Stratigraphic distribution and abundance data of ichnofossils in the succession.....7.3
- Fig. 7.3 Stratigraphic distribution of deep-sea and flysch ichnospecies through time.....7.9
- Fig. 7.4 Stratigraphic distribution of deep-sea and flysch ichnogenera through time.....7.10
- Fig. 7.5 Relative distribution of both deep-sea and facies-crossing ichnogenera through time...7.10
- Fig. 8.1 Schematic diagram showing the 15 collected ichnogenera and their ichnofacies ranges...8.46
- Fig. 8.2 Environmental distribution and abundance data of the ichnospecies from the study area.....8.47
- Fig. 8.3 Ethological groupings and possible producing

organisms for the ichnogenera recorded  
from the study area.....8.49

**LIST OF TABLES**

**After page**

Table 4.1	Runs test results.....	4.9
Table 5.1	Framework grain mode parameters of sandstones from the southern Welsh Basin.....	5.4
Table 5.2	Representative chemical analyses of sandstones and mudstones from the southern Welsh Basin.....	5.7
Table 5.3	Mean SiO <sub>2</sub> wt% and K <sub>2</sub> O/Na <sub>2</sub> O values of sandstone-dominated formations from the southern Welsh Basin.....	5.10
Table 5.4	Trace element geochemical parameters for mudstone-dominated formations from the southern Welsh Basin.....	5.12

**LIST OF PLATES**

**After page**

Plate 2.1	Sediment facies from the study area.....	2.2
Plate 2.2	Sediment facies from the study area.....	2.5
Plate 4.1	Depositional environments in the study area.....	4.11
Plate 4.2	Depositional environments in the study area.....	4.13
Plate 6.1	BSEM images of pyrite framboids.....	6.6
Plate 8.1	Ichnofossils from the Welsh Basin.....	8.9
Plate 8.2	Ichnofossils from the Welsh Basin.....	8.24
Plate 8.3	Ichnofossils from the Welsh Basin.....	8.36
Plate 9.1	<u>Desmograptus</u> <u>fuchsi</u> .....	9.5

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## CHAPTER ONE - HISTORICAL PERSPECTIVE AND BASIN STRATIGRAPHY

### 1.1 History of Studies

The coastal cliffs of Ceredigion and Preseli between Dinas Head and New Quay, provide the most extensive sections through the predominantly Ordovician and early Silurian rocks of mid-Wales (Fig. 1.1). Despite this, the Llandeilo-Llandovery coastal succession between Dinas Head and New Quay has received little attention from geologists in the last 25 years (Fig. 1.2). Lowman (1977) and Lowman & Bloxam (1981) examined the Fishguard Volcanic Group which stratigraphically underlies the oldest rocks in the study area ("Hendre Shales" of Lowman & Bloxam 1981) and merely noted the presence of the overlying mudstones. Hendricks (1926) mapped the coast around the village of Llangranog, this work being later updated by Anketell (1963). Both of these studies are essentially stratigraphic in nature. Craig (1985) completed a structural map of the area from Gwbert to Tresaith and linked it with the structural maps provided by Anketell (1963) for the Llangranog area and Anketell & Lovell (1976) for the New Quay area.

Smith (1956) and Wood & Smith (1959) undertook a major study of the Aberystwyth Grits Formation in the northern part of the region. The only sedimentological papers on this area are by James (1975) and Allen (1981) on the sediments of the Poppit Sands Formation, in the Cardigan area.

### 1.2 Aims of the present study

Taking account of the nature of the previous geological work on the study region it was decided to concentrate on a number of geographically discrete sub-areas and topics in order to provide the basis for an overview of the entire succession. The main areas of scientific investigation undertaken were:

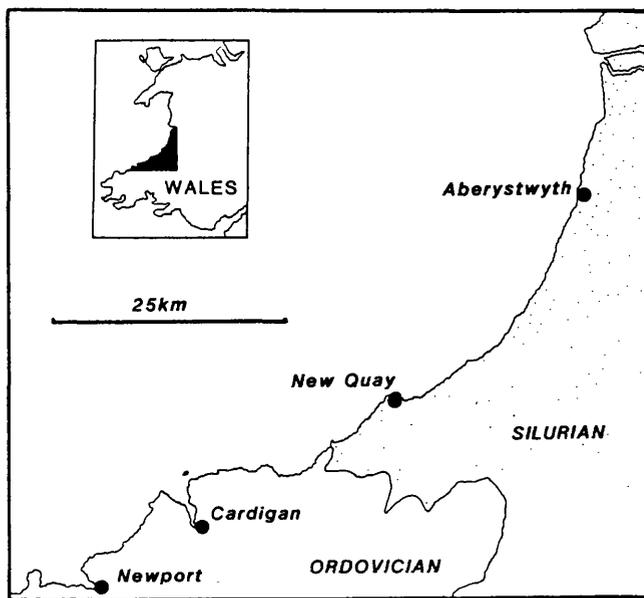


Figure 1.1 Map of Wales showing study area.

## 1.2

(1) Study of the southern part of the area, from Dinas Head (SN 008413) to Cardigan (SN 179461). This coastline had not been examined in any previous studies. This is partly due to its largely inaccessible nature. This area was mapped at a scale of 1:10,000. This involved examination both of road and stream sections and also a photo-survey of the coastline by boat. All major outcrops were logged and graptolites were collected where possible. Four formations (Parrog, Newport Sands, Ceibwr and Poppit Sands) were designated in the area.

(2) Study of the northerly parts of the succession, between Gwbert (SN 158502) and New Quay (SN 388605), particularly those formations designated by Craig (1985), Hendricks (1926), Anketell (1963) and Smith (1956). This has resulted in the recognition of recurring lithologies within the area.

(3) A significant part of this thesis is the petrographic and geochemical analysis of the sandstones and mudstones from the coastal succession to determine their provenance. The lack of previous data had been noted by Bassett (1984).

(4) The variety of trace fossils found in the succession are described together with the possible controls on their distribution.

### 1.2 Basin stratigraphy

The possible tectonic setting of the Welsh depositional Basin has been reviewed recently by a number of authors, notably Holland *et al.* (1979), Bassett (1984), Woodcock (1984a) and Siveter *et al.* (1989). In order to understand the geological setting of the Lower Palaeozoic Welsh Basin and to appreciate fully the importance of the geology of the study area, a summary of the various plate tectonic models is presented herein.

The area of Wales and its contiguous margins across the Welsh Borderland and English Midlands is generally interpreted as having been the site of a fault-bounded marginal depositional

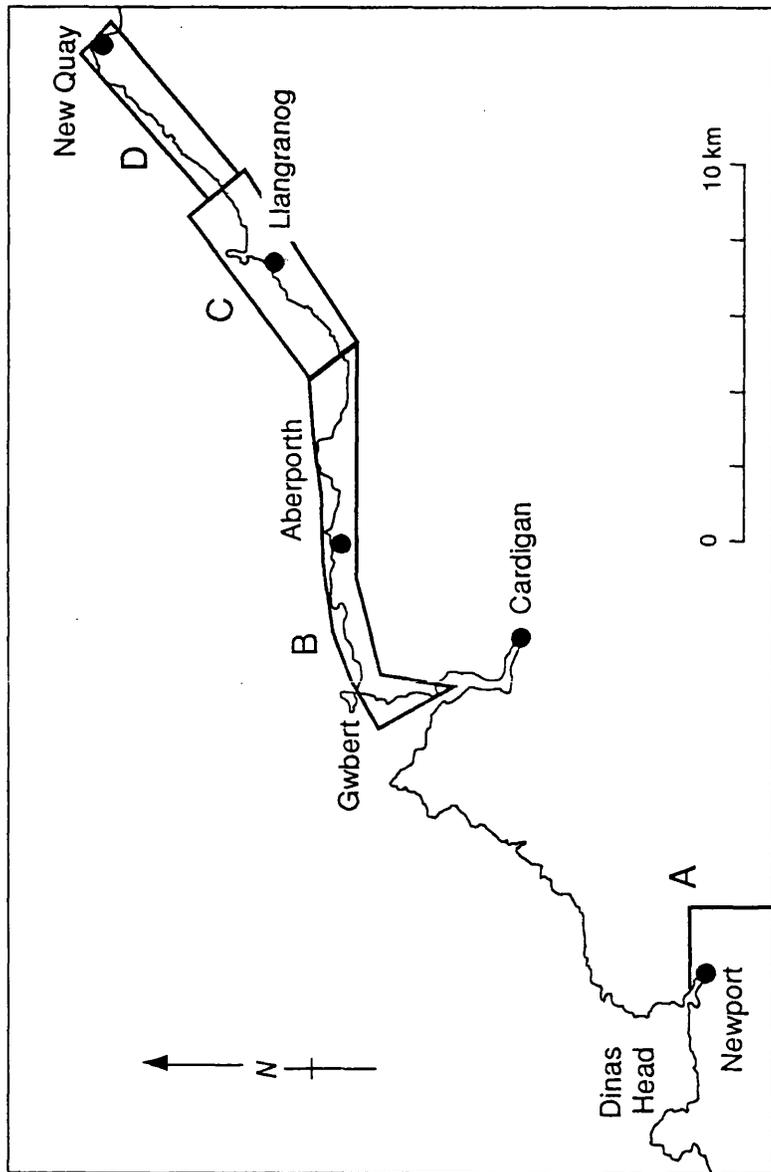


Figure 1.2 Map of study area showing previously published work: A, (Lowman & Bloxam 1981); B, (Craig 1985); C, (Hendricks 1926; Anketell 1963); D, (Smith 1956; Wood & Smith 1959; Anketell & Lovell 1976).

basin positioned on the southern side of the Iapetus Ocean from late Precambrian through early Palaeozoic times (Fig. 1.3). The axis of the Lower Palaeozoic Welsh Basin trends approximately NE-SW through central Wales. An ensialic back-arc basin setting is widely favoured for most of Ordovician time (e.g. Kokelaar *et al.* 1984; Leat & Thorpe 1986) but there is no agreement on an appropriate model for the period (post-Caradoc) which followed the cessation of major volcanic activity. Models proposed include a passive margin setting (Davies & Cave 1976), a fore-arc basin (Okada & Smith 1980) and a strike slip continental borderland (Woodcock 1984a).

The Lower Palaeozoic Welsh Basin was situated on the southern margin of the Iapetus Ocean, which separated the microcontinent of Avalonia (incorporating the southern parts of Ireland and Britain) from Laurentia (which incorporated northern Britain, northern Ireland, Scotland, Newfoundland and parts of the eastern seaboard of North America) and Baltica (incorporating much of present day Scandinavia). The Iapetus Ocean was formed around 1000 Ma ago as a result of intracontinental rifting, followed by continental fragmentation in late Precambrian times (800-750 Ma) (Kumpulainen & Nystuen 1986).

Some current models suggest closure of the Iapetus Ocean from Silurian to early Devonian times. Prior to this the ocean was continuous and mostly wide between Laurentia to the north and Baltica, Gondwana and Avalonia to the south (e.g. Barker & Gayer 1985; McKerrow & Cocks 1986). Pickering *et al.* (1988) and Pickering (1989) have noted that the available seismic, stratigraphical, structural, igneous, faunal, palaeomagnetic and sedimentological information from Britain, Scandinavia and Newfoundland suggest that possibly by the early Silurian, the Iapetus Ocean had closed, at least in part, with the consumption of intervening oceanic crust. Seaways, however, remained important, at least in the British sequences, at least until the late Silurian (Pickering *et al.* 1988). The early Palaeozoic orogeny in Wales culminated during the closure of the Iapetus Ocean.

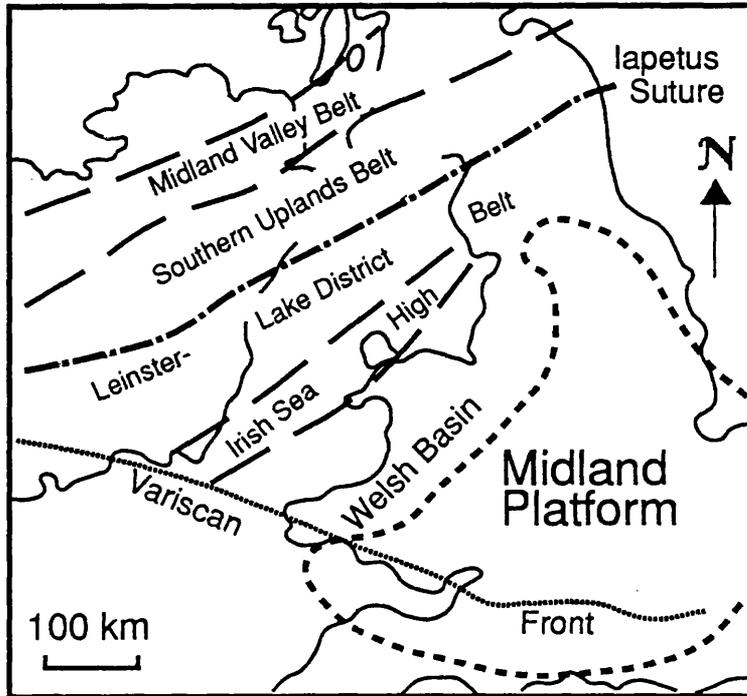


Figure 1.3 Regional and tectonic framework of the Welsh Basin (modified after Ziegler 1984).

The Welsh Basin was tectonically active throughout its early Palaeozoic history (Woodcock 1984a). Deformation of the Lower Palaeozoic rocks of the Welsh Basin involved both dip-slip and strike-slip movements. Several of the major fault complexes, for example the Pontesford and Church Stretton lineaments (Woodcock & Gibbons 1988) demonstrate basement control by the relatively young Precambrian basement. Furthermore a number of the major fault complexes (Fig. 1.4), particularly in the north-western and southeastern margins of the Wales-Welsh Borderland depositional basin, have been interpreted as possible terrane boundaries, based on the lack of stratigraphic continuity across them (Woodcock & Gibbons 1988).

The Menai Strait Fault System, across Anglesey and the coast of North Wales, is the best contender for a terrane boundary. Here the Monian terranes docked against the incipient Welsh Basin sometime during the late Precambrian-early Cambrian as part of the Cadomian tectonic event (Woodcock & Gibbons 1988). The Malvern Lineament, along the Welsh Borderland, is a fault system that possibly represents another Precambrian terrane boundary. The Welsh Borderland Fault System may have resulted from later extensional tectonics during the main, Cambrian-Ordovician developmental phase of the depositional basin. However, post-Llandovery stratigraphical continuity across the Welsh Borderland Fault System, onto the Midland Platform, precludes its representing a post-Ashgill terrane boundary (Woodcock & Gibbons 1988; Siveter et al. 1989).

The Welsh Basin is underlain by late Precambrian age, calc-alkaline plutonic and igneous rocks and sediments which are interpreted as an ancient volcanic arc (Thorpe 1979; Watson & Dunning 1979). The geometry and location of the overlying fault systems was largely determined by basement fault patterns (Woodcock 1984a). In turn, the sites and patterns of both sedimentation and volcanic activity were affected by the nature and distribution of the major structural displacements (e.g. Kokelaar et al. 1984; Fitches & Campbell 1987; Kokelaar 1988). Between the Tremadoc and the middle Ordovician the geochemistry and style of volcanic

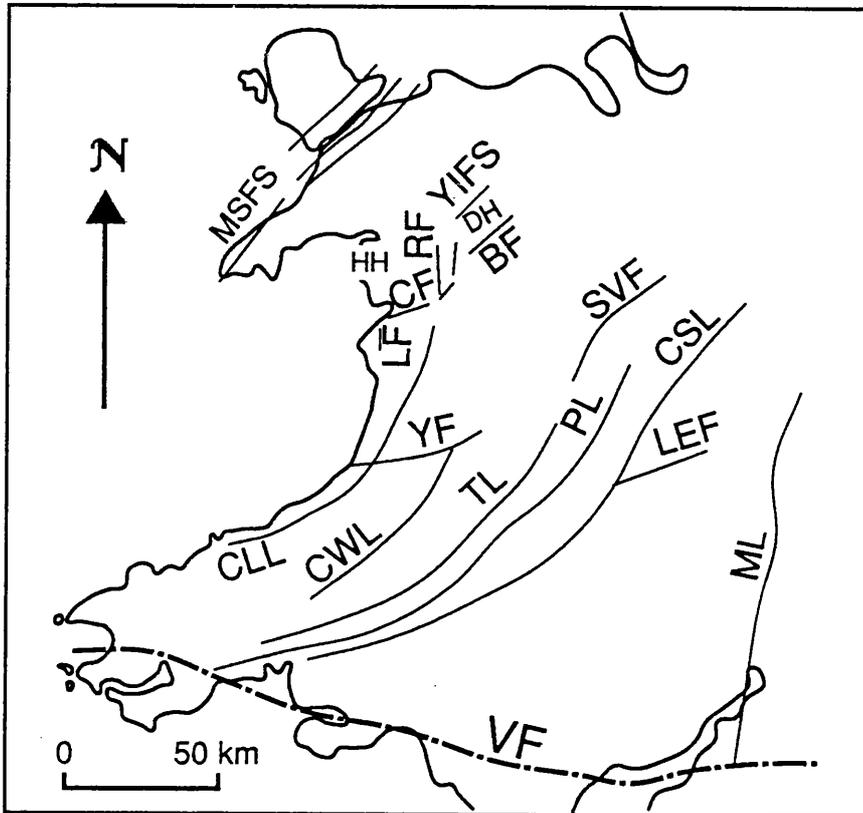


Figure 1.4 The "Variscan Front" and structures inferred to have been active during the early Palaeozoic history of the Welsh Basin: BF, Bala Fault (James & James 1969; Fitches & Campbell 1987); CLL, Corris-Llangranog Lineament (James & James 1969; James 1972; Craig 1985; Craig 1987); CF, Ceunant Fault (Fitches & Campbell 1987); CSL, Church Stretton Lineament (Holland & Lawson 1963); CWL, Central Wales Lineament (Smith 1987a); DH, Derwen Horst (Fitches & Campbell 1987); HH, Harlech Horst (Fitches & Campbell 1987); LEF, Leinthall Earls Fault (Lawson 1973); LF, Llanegryn Fault (Fitches & Campbell 1987); ML, Malvern Line (Hurst *et al.* 1978); MSFS, Menai Straits Fault System (Kokelaar *et al.* 1984); PL, Pontesford Lineament (Woodcock 1984b); RF, Rhobell Fracture (Kokelaar *et al.* 1984; Fitches & Campbell 1987); SVF, Severn Valley Fault (James & James 1969); TL, Twyi Lineament (Smallwood 1986); YIFS, Ysppyty Ifan Fault System (Fitches & Campbell 1987); YF, Ystwyth Fault (James & James 1969).

eruptions changed, signalling a transition from a volcanic arc to a more marginal basin setting (Kokelaar 1988).

The Welsh Basin was defined on the east and south by a slope rising to the shelf-sea areas of the Welsh Borderland and south Wales. To the north the contemporary basin margin is less well defined, but probably extended northwestwards towards the North Wales coast. The main part of the Welsh Basin persisted throughout the early Palaeozoic, and was characterised by prolonged deposition of dark graptolitic shales with periodic influxes of submarine fan sediments. The basin may have been many hundreds of metres deep (Cave & Hains 1986) particularly in its southern reaches (Woodcock 1984a).

Ashgill and early Llandovery turbidite systems in the Welsh Basin were sourced mainly from the eastern lateral basin margin with only minor axial supply from the south (Cave 1979). The importance of the lateral source decreased during the late Llandovery transgression of the Midland Platform while the southern source became far more important supplying the sediment for the late Llandovery and Wenlock turbidite systems (Dimberline 1987; Smith 1987b).

## 1.2 Thesis Organisation

This thesis is arranged into nine chapters, each styled and referenced for a particular journal.

Chapter one presents the aims of the project together with the organisation of the thesis and the methodology used.

Chapter two describes the main lithofacies recognised from the study area, together with possible depositional mechanisms.

Chapter three is a more detailed examination of the sediments of the area. The chapter is divided into two parts, the first of which is devoted to the area between Dinas Head and

Cardigan, while the second part details the area from Cardigan to New Quay. The sediments of each formation are described and representative stratigraphic logs are presented for each formation. A stratigraphic map showing the distribution of the formations through the succession is presented.

Chapter four (submitted to Geological Journal) arranges the sediments of the area into facies associations, the distribution and controls of which are outlined. Based on palaeocurrent analysis and the inferred depositional environments the palaeogeography of the succession, within the Welsh Basin, is presented for Llandeilo-Llandovery times. The controls on sedimentation are discussed and a modern analogue is presented for the Lower Palaeozoic Welsh Basin. Part of this chapter has been published in the Journal of the Geological Society of London (McCann & Pickering 1989, in press).

Chapter five (accepted for publication in a Special Publication of the Geological Society of London: Developments in Sedimentary Provenance) is an analysis of the provenance of the sediments of the study area. Sandstone provenance was determined by framework grain analysis together with whole-rock major element geochemistry. These data were integrated with the major and trace element geochemical data of the mudstones and then related to probable source areas.

Chapter six examines in detail the fine-grained sediments of the succession, particularly their optical and geochemical properties. The chapter also evaluates a model (Dimberline 1987) for hemipelagic deposition.

Chapter seven (provisionally accepted by Lethaia) analyses the distribution of the trace fossil assemblages through the succession.

Chapter eight (submitted to Palaeontology) systematically describes the collected ichnospecies from the succession.

Chapter nine (McCann 1989) is a detailed taxonomic review of the ichnogenus Desmograpton fuchsi. This is the first record of this characteristic deep-water trace fossil both from the Palaeozoic and from Britain.

The major conclusions of the study, together with suggestions for further research, are presented in Chapter ten.

## 1.4 Methods and techniques

### 1.4.1 Measurement

The metric system is used throughout. Sediment grain sizes were determined using the Wentworth (1967) scale while bed thicknesses were classified after Ingram (1954). Structural and palaeocurrent information were recorded using a Silva compass clinometer.

### 1.4.2 Mapping

Mapping was done on 1:10,000 scale Ordnance Survey sheets for the Dinas Head to Cardigan area. The coastal succession provides the best outcrop although access is severely restricted. Much of the mapping of the area was undertaken from boats with samples being collected from the larger beaches. Further data were collected from the Pembrokeshire Coastal Footpath. Inland exposure is rare and is generally of very poor quality. Where present it is confined to small roadside sections.

The dominant structure of the Dinas Head to Cardigan area comprises a series of folds plunging gently to the northeast. This is similar to the structure of the area north of Cardigan as outlined by Craig (1985), Hendricks (1926) and Anketell (1963), with the fold axes paralleling the major tectonic lineament trends. The fold limbs commonly comprise a complex of smaller parasitic folds which may be almost isoclinal.

### 1.4.3 Section Logging

Logging of sedimentary successions is the major technique used to characterise the sediments present. Logging was carried out at the centimetre scale for all sandstone-dominated sections and at a millimetre scale for those dominated by mudstones.

### 1.4.4 Laboratory techniques

Bulk rock major and trace element geochemistry was determined using a Phillips X-ray generator. Pressed powder pellets were prepared from samples which were crushed and then finely ground in a *tama*. The results were used both for determination of provenance of sandstones and mudstones and also for more detailed examination of the finer-grained fractions. X-ray diffraction analyses, using a Phillips diffractometer with associated electronic counting apparatus and Ni-filtered Cu-K-alpha radiation, were determined for mudstone turbidites and hemipelagic sediments in an attempt to aid their classification. Total organic carbon (TOC) contents were measured using a Carbon-Sulphur Determinator (CS-125).

Polished blocks were used to examine the facies types and sedimentary structures of fine-grained sediments in detail. Backscattered scanning electron microscopy linked with qualitative energy dispersive X-ray microanalysis was used for information on mineralogy and fabric at high resolution.

Point-counting was carried out by means of an automatic point counter using a Swift binocular microscope. A number of thin sections were stained for K-feldspar following the procedure outlined in Deer et al. (1966).

### 1.4.5 Material repository

Representative collections of thin sections, graptolites and

ichnofossils have been deposited in the National Museum of Wales, Cardiff. The collection number is NMW89.13G. Where appropriate these are indicated in the figure caption.

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**CHAPTER TWO - SEDIMENTARY FACIES****2.1 Introduction**

The rocks in the area have been subdivided into a number of sedimentary facies. In this section, the facies are defined primarily using the scheme and process interpretations as outlined in Pickering et al. (1986) and Stow & Piper (1984).

**2.2 Disorganised gravels, gravelly muds and pebbly sands**

This facies is the most coarse-grained sediment in the study area, with greater than 5% pebble grade or coarser material. They are volumetrically the least important, cropping out in a limited number of areas. Bed thicknesses are extremely variable and may pinch out laterally.

**2.2.1 Disorganised gravelly muds (A1.3) (Plate 2.1a)**

This facies consists of structureless gravelly muds with 50-95% mud grade sediment. The matrix-supported clasts are freely distributed throughout the deposit with no evidence of grading or internal organisation. Clasts are subrounded to subangular and range in size from 1.0-23.0 cm. The clasts consist primarily of small quartz pebbles, sandstone, siltstone and mudstones. Some fossil fragments, for example crinoid stems, are also included.

Deposition was either from freezing of a cohesive debris flow or a slide deposit.

**2.2.2 Disorganised pebbly sands (A1.4) (Plate 2.1b)**

This facies is a structureless sandy deposit which contains a large percentage of bioclastic material. The clasts include

## 2.2

trilobite and brachiopod fragments together with pelmatozoan debris. Scouring and loading are common. Mudstone rip-up clasts are also common, and locally the deposit may be termed a mud-flake breccia. There is no real evidence of vertical grading although clast concentration may be laterally variable.

Deposition was from high-concentration turbidity currents with the material being transported long distances.

## 2.3 Sandstones

Sandstone deposits are described as those with greater than 80% sand grade fragments and less than 5% pebble grade clasts. Volumetrically these deposits are important in a number of formations in the area, for example the Poppit Sands and Aberystwyth Grits formations. Bed thicknesses vary between 6.0-200.0 cm and are both continuous and discontinuous over the extent of outcrop.

### 2.3.1 Thick- to medium-bedded disorganised sandstones (B1.1) (Plate 2.1c)

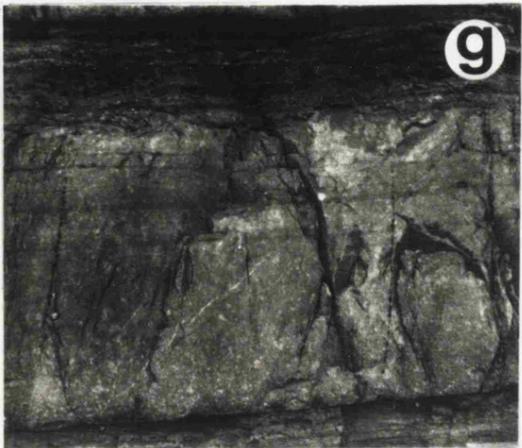
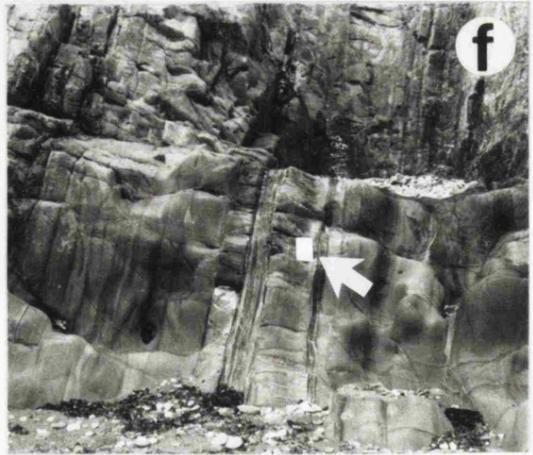
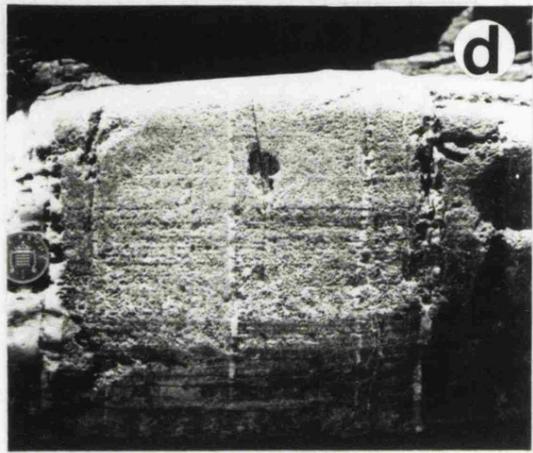
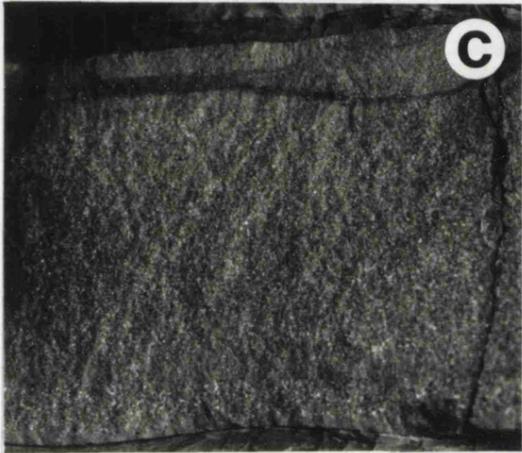
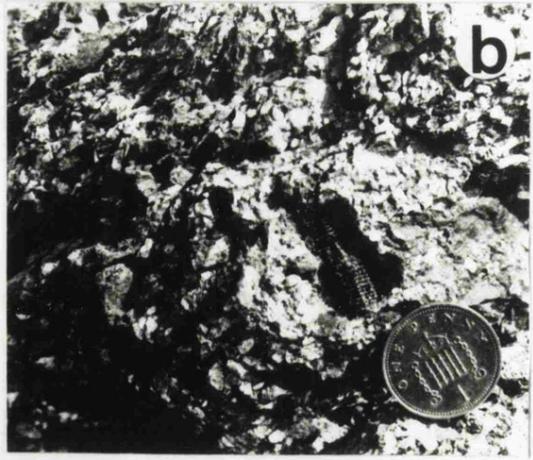
This facies comprises thick- to medium-bedded sands lacking grading and typically showing sharp, flat bounding surfaces. Coarse-tail grading, with granules concentrated in a less than 1.0 cm thick basal layer, may occur. Dish structures have been reported from examples of this facies at Poppit Sands (Allen 1981).

Deposition was probably from a high-concentration turbidity current by freezing of a dense cohesionless suspension. The presence of dish structures, however, would suggest some post-depositional fluidisation.

### 2.3.2 Thick to medium-bedded parallel stratified sandstones (B2.1) (Plate 2.1d)

PLATE 2.1

- a. Disorganised gravelly muds (A1.3) from the Poppit Sands Formation (Caradoc), Poppit Sands (SN 14604960). The photograph is 30.0 cm wide.
- b. Disorganised pebbly sands (A1.4) from the Poppit Sands Formation (Caradoc), Poppit Sands (SN 14604960). Coin (2.0 cm) for scale.
- c. Thick- to medium-bedded disorganised sandstone bed (B1.1) from the Aberystwyth Grits Formation (Upper Llandovery), New Quay (SN 38686041). The sandstone bed is 34.0 cm thick.
- d. Thick- to medium-bedded parallel laminated sandstone (B2.1) from the Poppit Sands Formation (Caradoc), Poppit Sands (SN 14604960). Coin (2.0 cm) for scale.
- e. Poorly sorted muddy sands (C1.1) from the Ceibwr Formation (Llandeilo-Caradoc), Ceibwr Bay (SN 10224510). Stripe on hammer (3.0 cm long) for scale.
- f. Very thick to thick sand-mud couplets (C2.1) from the Newport Sands Formation (Llandeilo), Newport Sands (SN 05445082). Notebook (20.0 cm long) for scale.
- g. Medium-bedded sand-mud couplets (C2.2) from the Poppit Sands Formation (Caradoc), Poppit Sands (SN 14604960). The sandstone bed is 18.0 cm thick.
- h. Thin-bedded sand-mud couplets (C2.3) from the Poppit Sands Formation (Caradoc), Poppit Sands (SN 14604960). Coin (2.0 cm) for scale.



## 2.3

The facies is typified by thick to medium-bedded, medium- to granule-grade sandstones with horizontal to near horizontal stratification either throughout or towards the top of the bed. Bed shape is commonly planar and amalgamation is common.

Lower bounding surfaces are generally sharp, although minor scouring with a resultant granule-grade lag, may be locally developed. Above this lies a thin, medium- to granule-grade, inversely graded layer, termed a traction carpet, which is, in turn, overlain by a medium- to coarse-grained division which grades and fines upwards. The uppermost part of the bed comprises parallel-laminated sands and possibly a cross-laminated division.

This facies was probably produced by the freezing of successively generated traction carpets at the base of a highly concentrated turbidity current. The more structureless divisions are the result of rapid grain-by-grain suspension fall-out or freezing of a thicker unsorted layer.

## 2.4 Sand-Mud Couplets

This facies grouping consists of sand-mud couplets which comprise 20-80% sand and less than 80% mud/silt grade. Bed shape is variable but is commonly sheet-like. Beds from this facies are best described and subdivided using the Bouma (1962) sequence. Amalgamation may be common in the thicker beds of the facies.

### 2.4.1 Poorly sorted muddy sands (Cl.1) (Plate 2.1e)

The facies comprises poorly-sorted, mud-rich (up to 80%) sands showing some evidence of normal grading. The beds are commonly normally graded in the coarser sand-grade portions. Bed bases are sharp with occasional well-developed bottom structures, for example loading and flame structures, while

## 2.4

the tops are graded and mud rich. Mudstone rip-up clasts may be present within the beds.

Deposition was from muddy high-concentration turbidity currents or fluid sand-mud debris flows and may have been rapid.

### 2.4.2 Very thick to thick-bedded sand-mud couplets (C2.1) (Plate 2.1f)

The facies consists of very thick to thick sand-mud couplets comprising medium- to coarse-grained sandstones which are normally graded. Bouma (1962) Tabc divisions are commonly well developed. Bottom structures such as flutes and load casts may be present.

Deposition was from high concentration turbidity currents.

### 2.4.3 Medium-bedded sand-mud couplets (C2.2) (Plate 2.1g)

The facies comprises medium-bedded sand-mud couplets with well developed normal grading and commonly Tbcd divisions although complete Ta-e sequences may be locally developed. The sandstones are fine- to coarse-grained and flute casts and load structures are locally developed.

Deposition was from high concentration turbidity currents.

### 2.4.4 Thin-bedded sand-mud couplets (C2.3) (Plate 2.1h)

The facies is typified by thin-bedded sand-mud couplets with well developed normal grading and commonly base-missing Bouma (1962) sequences, for example, Tbcde, Tcde, Tce and Tde. The sandstones are very fine- to fine-grained.

Deposition was from low concentration turbidity currents.

## 2.5 Silts, Muddy Silts and Silt-Mud Couplets

This facies contains those sediments which are dominantly silt and mud grade. The sediments comprise more than 80% mud grade, 40% or more of which is silt grade. A small proportion, commonly less than 20%, is of slightly coarser sand grade. Silt has therefore replaced sand as the dominant 'coarse' grain size.

### 2.5.1 Graded stratified silts (D2.1) (Plate 2.2a)

The facies comprises normally graded, stratified silts which are thin- to medium-bedded. Bases are commonly sharp and may be scoured. Bed tops are gradational. Internally, the beds may be classified according to the Bouma (1962) sequence. Typically beds show Tbce, Tcde and Tde divisions.

Deposition was from low-concentration turbidity currents.

### 2.5.2 Thick irregular silt and mud laminae (D2.2) (Plate 2.2b)

This facies consists of medium- to thick-bedded lenticular silt laminae in mud or thin irregular convolute silt laminae and lenses. Loading is locally present and in some cases may be extreme with the production of laminated pseudonodules and intervening mud flame structures.

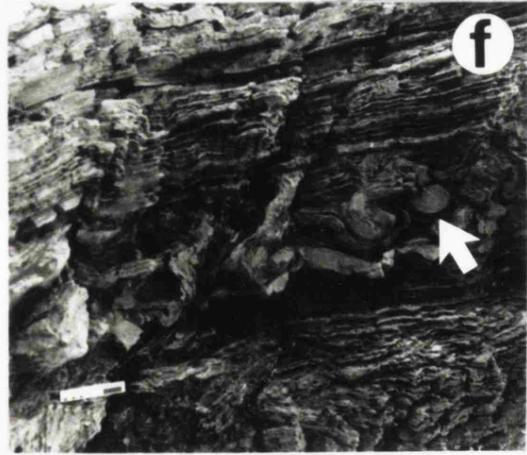
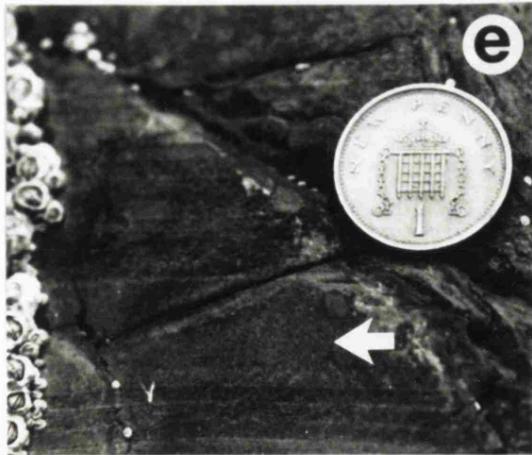
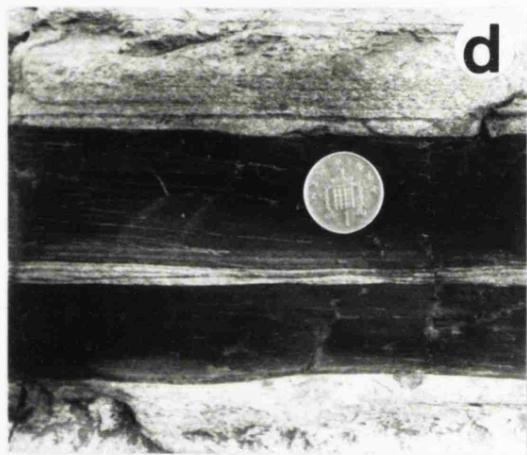
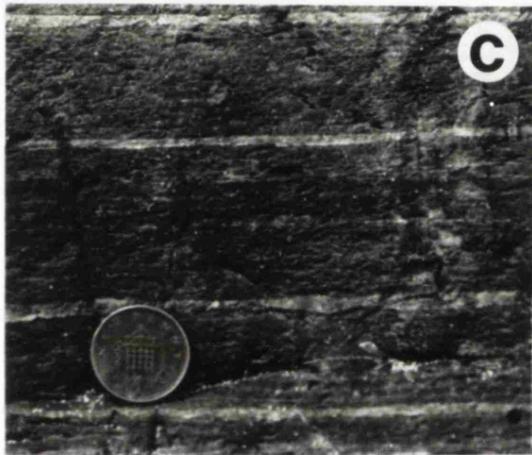
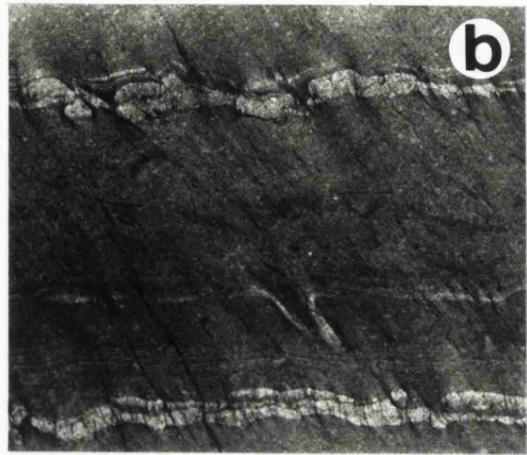
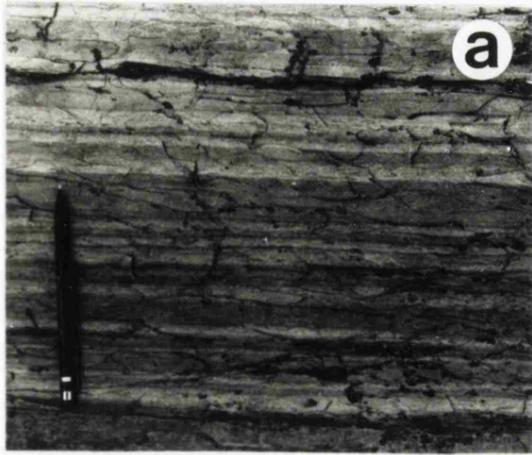
Deposition was rapid and from low concentration turbidity currents.

### 2.5.3 Thin regular silt and mud laminae (D2.3) (Plate 2.2c)

This facies is typified by thin- to medium-bedded, parallel to subparallel, silt laminae in mud. The laminae are commonly in graded laminated units and may be slightly lenticular.

PLATE 2.2

- a. Graded stratified siltstone beds (D2.1) from the Gwbert Formation (?Caradoc), Gwbert (SN 16185121). Pencil (14.0 cm) for scale.
- b. Thick irregular silt and mud laminae (D2.2) from the Gwbert Formation (?Caradoc), Gwbert (SN 16185121). The photograph is 42.0 cm wide.
- c. Thin regular silt and mud laminae (D2.3) from the Gaeroglwyd Formation (Lower Llandoverly), Traeth-yr-Ynys (SN 31905495). Coin (2.0 cm) for scale.
- d. Structureless muds (E1.1) from the Poppit Sands Formation (Caradoc), Poppit Sands (SN 14604960). Coin (2.0 cm) for scale.
- e. Graded muds (E2.1) (arrowed) and laminated muds and clays (E2.2) from the Gaeroglwyd Formation (Lower Llandoverly), Traeth-yr-Ynys (SN 31905495). Coin (2.0 cm) for scale.
- f. Contorted/disturbed strata (F2.1) and balled strata (F2.2) (arrowed) from the Llangranog Formation (Ashgill), Llangranog (SN 31095426). Scale bar is 15.0 cm long.



## 2.6

Deposition was relatively slow and was from low-concentration turbidity currents.

### 2.6 Muds and Clays

Muds and clays locally predominate within the study area. They contain more than 95% mud grade sediment of which less than 40% is of silt grade. A small proportion, less than 5%, of sand grade material may be locally present.

#### 2.6.1 Structureless muds (E1.1) (Plate 2.2d)

This facies comprises structureless muds and clays commonly occurring in thick sections where bedding is poorly defined or absent. The sediments are mostly devoid of distinct sedimentary structures.

The facies was possibly produced as a result of relatively rapid deposition. It may also have resulted from both the ponding of thick diluted turbidity currents, and, or, hemipelagic settling.

#### 2.6.2 Graded muds (E2.1) (Plate 2.2e)

The facies consists of well-bedded colour graded muds often with very thin silt laminae at the base. Two types of mud occur: dark grey and a light grey mudstone. Boundaries between them are gradational. Bioturbation is generally present and is confined to the light grey mudstone layers. Within these units it may be concentrated towards the top or bottom. It is rarely evenly distributed throughout the bed.

Deposition was from low and high concentration turbidity currents.

### 2.6.3 Laminated muds and clays (E2.2) (Plate 2.2e)

The facies comprises finely-laminated organic-rich muds and clays. These are commonly found in mud-rich formations where they may be locally dominant. Individual beds range from centimetres to decimetres in thickness and may be interbedded with graded mudstone turbidites (Facies E2.1). The laminae are alternately organic-rich and mud-rich and thus show both compositional and colour variations. Bioturbation is extremely rare and has only been observed in the interbedded graded mudstones.

Deposition was by grain-by-grain or aggregate settling and from low concentration turbidity currents. Anoxic bottom waters favour the preservation of high concentrations of organic matter.

## 2.7 Chaotic Deposits

This facies includes chaotic mixtures of sediments and rock types that, for the most part, have been subjected to large-scale downslope mass movements.

### 2.7.1 Contorted/disturbed strata (F2.1) (Plate 2.2f)

This facies consists of coherent folded and contorted strata. The facies is lithologically diverse comprising thick- and thin-bedded sandstones, siltstones and mudstones. It was produced mainly by gravity-induced sediment sliding and slumping in which the shear strength of the sediment was exceeded.

### 2.7.2 Dislocated and balled strata (F2.2) (Plate 2.2f)

This facies is typified by dislocated and balled strata in a jumble of fragments. They are commonly concentrated at the tops of sandstone turbidite beds although they may also form

## 2.8

isolate beds. The facies is commonly intraformational. It was produced mainly as a result of in-situ liquefaction and fluidization.

## 2.8 References

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## CHAPTER THREE - STRATIGRAPHY AND SEDIMENTOLOGY OF THE DINAS HEAD TO NEW QUAY AREA

### Abstract

This chapter outlines the stratigraphy of the study area. Each of the formations recognised in this study is described in detail and representative logged sections are presented. Sedimentological analysis of the study area is divided into two parts: Part One examines the study area from Dinas Head to Cardigan, herein described for the first time. Part Two describes the sediments outcropping along the coast from Cardigan to New Quay. Palaeoenvironmental interpretations and palaeogeographic reconstructions are included in Chapter Four.

### Part One

#### 3.1 Introduction

The Dinas Head to Cardigan area is in the southern part of the study area. It's rocks are the oldest in the study area (cf. Latter 1925), but their exact age and age relationships within the area have been poorly understood before the present study.

The region was mapped at a scale of 1:10,000 and this provided a suitable framework for the delineation of the four lithostratigraphic units which are described below. The main roads (A487, B4582) were taken as an arbitrary inland boundary of the mapping area. West of these roads the entire geology to the coast was mapped. Inland it is difficult to correlate the lithological units and lateral changes in sedimentation with any degree of certainty as exposure is very poor. The discussion of the area is restricted to specific points, particularly the distribution of the

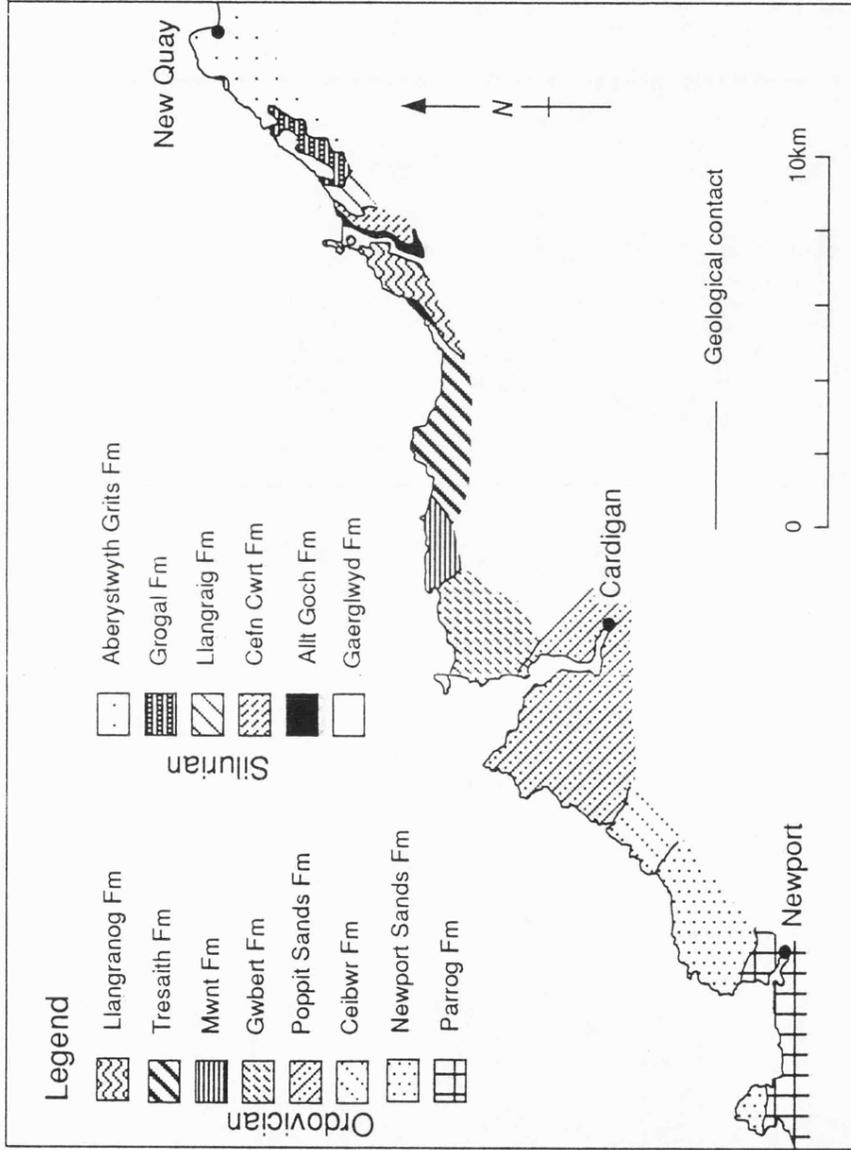


Figure 3.1 Stratigraphic map of the Dinas Head to New Quay area. Stratigraphy derived from the following sources: Newport-Cardigan (present study); Cardigan-New Quay (Hendricks 1926; Wood & Smith 1959; Anketell 1963; Craig 1985).

sedimentary facies and the age and stratigraphic relationships within the region. A more general discussion of the geology of the area is presented in chapter four.

### 3.2 Previous Work

The rocks of the region were placed stratigraphically below the Old Red Sandstone in the earliest geological maps of the area (Challinor 1969). The BGS geological map of the Newport region (Sheet 40 - May 1857), however, classified the rocks on the eastern side of Newport Bay as Llandeilo in age. Keeping (1881) said that the rocks of the Newport and Cardigan area were of "Middle Bala" or Caradoc age. Later he suggested a Llandeilo or Bala age based on some poorly preserved graptolites (Keeping 1882). Lapworth (in Keeping 1882) also examined the graptolitic finds but could not say if they were of Arenig, Llandeilo or Lower Llandovery age.

Cowper Reed (1895) noted the similarity between the rocks of Dinas Head and Newport, although he figured the rocks as Llandovery in age while those on the south side of Newport Bay were shown as upper Llandeilo and Bala (Caradoc-Ashgill). Matley (1897) also noted the similarities between the rocks cropping out on Dinas Head and those to the north of Newport and classed them as of Bala (Caradoc-Ashgill) age. Latter (1925), in a short paper on the age of the rocks in the Newport to Cardigan area, suggested a gradual younging to the northeast. Jones (1955) considered that the area from Fishguard to Cardigan was Caradoc in age. Myers (1950) collected about twenty-five graptolite specimens from the eastern side of Newport Bay. Subsequent identification of the specimens indicated a horizon somewhere about the Glyptograptus teretiusculus - Nemagraptus gracilis boundary or a little higher thus confirming the age of the base of the succession in the area as Llandeilo.

To the north at Poppit Sands (SN 146496), there is a sedimentary succession which has been interpreted as turbidites of Caradoc age (James 1975). The sequence also

contains the first published occurrence of dish structures in the British Isles (Allen 1981), although these sedimentary structures have been widely reported from turbiditic sequences elsewhere in the world (Lowe 1975).

### 3.3 Stratigraphy

The rocks of the Dinas Head to Cardigan area are generally equivalent in age to the Corris - Cader Idris succession in the north and the rocks of the Aberiddy area to the south (Williams et al. 1972). The lithostratigraphic divisions in these adjacent areas, however, has not been used, as undoubtedly there is considerable lateral variation. Four new formations are proposed for the Dinas Head to Cardigan area (Fig. 3.1).

#### 3.3.1 Parrog Formation - Type locality: Parrog (SN 04553980)

The Parrog Formation comprises interbedded mudstones and siltstones with minor intercalated thin beds of medium- to fine-grained sandstone. It is well exposed along the coast from Cwm-yr-Eglwys (SN 016401) to Parrog (SN 045398). The succession is largely unfossiliferous and has a maximum thickness of 250m. The base is not exposed within the study area. The top is gradational with the overlying coarser, and lithologically more diverse, Newport Sands Formation, the contact coinciding with the increasing dominance of sandstone beds (SN 05404082). Rare Planolites ichnosp. are the only fossils found in the formation.

The mudstones are light grey and dark to blue grey, weathering to a buff, tan or reddish colour, and may be structureless, convolute laminated or parallel laminated. Early concretionary structures and disseminated pyrite may occur locally. X-ray diffraction analysis of the concretions reveals that they are rich in apatite which, according to Smith (1987), is a feature of early diagenesis. An unusual feature is the presence of lighter coloured bands within the



### 3.4

mudstones. These bands have an appearance initially consistent with bedding but on closer examination is seen to be a weathering product. The mudstones are very paper-like in character.

Siltstones are locally interbedded with the mudstones. The siltstones are light grey in colour and range from 1.0-30.0 mm in thickness. They are commonly lenticular and can be laterally persistent over distances of up to 10.0 m. They are internally structureless and parallel to cross-laminated.

Sandstone beds are rare but occur locally (e.g. Aber Fforest (SN 026396)). They are sharp based and laterally continuous over a scale of metres. They range from 2.0-3.0 cm in thickness and may have asymmetrically, or even bi-directional, rippled tops. Both the siltstone and sandstone beds may weather to a rust brown or orange colour.

Parrog Formation - Type locality: Section 1 (SN 04553980)

The section is located due west of Parrog, adjacent to the boat house, and is approximately 15.0 m thick of which 3.45 m were logged in detail (Fig. 3.2, 3.3). The section is dominated by light- and dark-grey interbedded mudstones as described above.

Facies types recognised at this locality include (C2.3), D2.1, D2.2, D2.3, E1.1 and E2.2. Facies without brackets are common while those within brackets are rare.

Mudstones are either structureless and laminated with the paper-like texture typical of this formation. They are unfossiliferous except for a single specimen of Planolites ichnosp. which was found adjacent to the section. Early diagenetic concretions, rich in apatite, are relatively abundant.

Siltstone beds are both laterally continuous or discontinuous over distances of metres, and may grade laterally from a

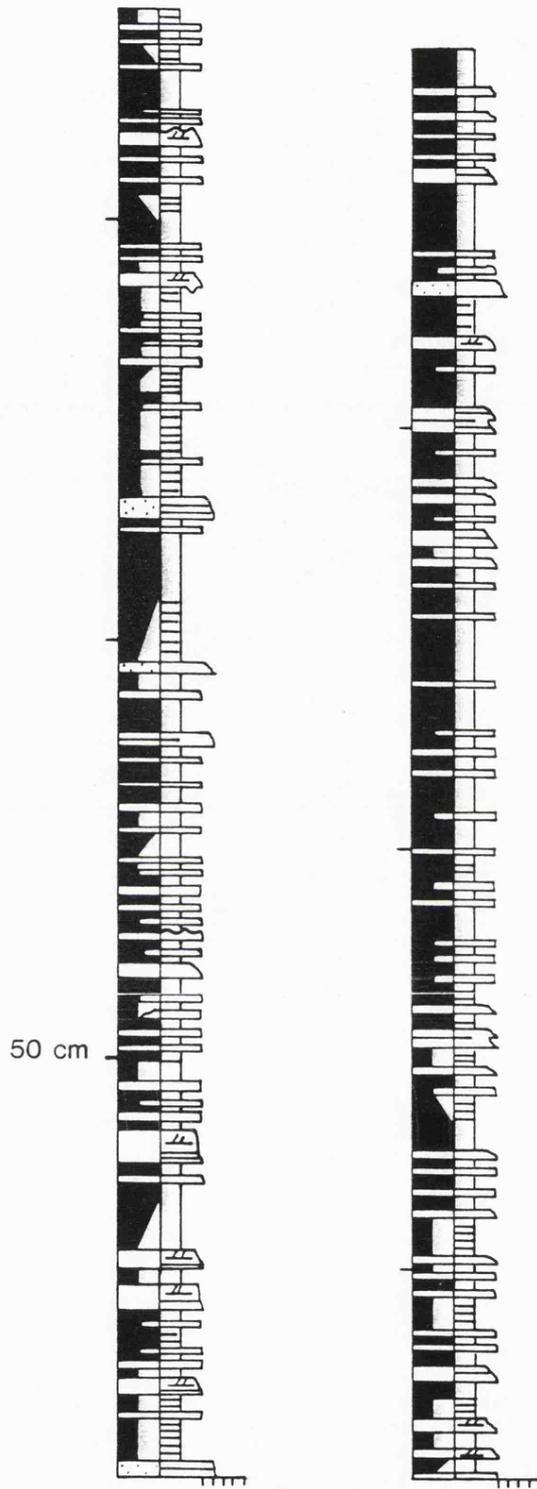


Figure 3.3 Measured section of Parrog Formation, Parrog.

### 3.5

single bed to a number of thinner beds. Bases may have load structures and there are rare ball and pillow structures. Most bases, however, are sharp. Internally, the siltstones display Tabe, Tbcde, Tbce, Tce and Tde divisions. Rare bed amalgamations have been noted.

Sandstones are rare, medium- to fine-grained and parallel laminated (Tde). An example of bidirectional rippling was noted adjacent to the section but this was the only occurrence in the Formation.

#### 3.3.2 Newport Sands Formation - Type locality: Newport Sands (SN 05454082)

The Formation consists mainly of fine- to coarse-grained sandstones which are either laterally persistent or scoured at a scale of tens of metres. Interbedded siltstones and mudstones are locally abundant. The Formation is well exposed along the coast from Newport Sands (SN 054408) to Carreg Yspar (SN 100450) although access is frequently restricted. The base of the formation at Newport Sands coincides with the marked decrease in the sandstone/mudstone ratio (SN 05404082) while the top of the formation is gradational with the overlying Ceibwr Formation. The upper contact is exposed on Traeth Cell-Howel (SN 088438)

The Newport Sands Formation contains a suite of graptolites, ichnofossils (Chondrites ichnospp., Palaeophycus ichnosp. and Planolites ichnosp.) and rare pelmatozoan debris. The graptolites include Climacograptus sp., Diplograptus ?multidens Elles & Wood, Glyptograptus siccatus Elles & Wood, G. cf. siccatus Elles & Wood, G. teretiusculus aff. euglyphus Lapworth, Glyptograptus sp. and Pseudoclimacograptus sp. (Dr R.B. Rickards pers. comm.; Collection site - SN 05354111). The presence of Glyptograptus siccatus Elles & Wood suggests that the assemblage belongs to the Nemagraptus gracilis Zone (Llandeilo). The Newport Sands Formation is up to 700 m thick but appears thicker due to structural repetition.

### 3.6

The light to dark blue-grey mudstones show no apparent structure although evidence of parallel lamination may be seen locally. They contain pyrite both in disseminated grains and in small aggregates. The mudstones commonly weather to an orange-brown colour.

Rare light grey siltstone bands up to 2.0 cm in thickness are interbedded with the mudstones. The siltstone beds have a sharp base and top and vary from laterally continuous to discontinuous over distances of 10's of metres.

Light to dark grey, fine- to medium-grained sandstones range in thickness from 1.0 - 33.0 cm. They are laterally persistent over the extent of outcrop exposure although the thinner and finer-grained sandstones are lenticular and laterally discontinuous. Bases are essentially planar and tops are rippled.

Coarse-grained sandstones occur as washouts, at the bases of finer-grained sandstones, and also as the basal (Ta) division of some amalgamated units. Grain size can be granule size at the base of some of these beds. The amalgamated units are up to 2.0 m in thickness and laterally continuous. Mudstone rip-up-clasts occur at the base, and more rarely the top, of some of the coarser sandstone beds. Some of the sandstones in the area weather to a characteristic yellow/rust colour not noted elsewhere in the study area.

Newport Sands Formation - Type locality: Section 1 (SN 05445082)

A coastal section, adjacent to Newport Sands, of which 26.0 m has been logged in detail (Fig. 3.4). There is extensive folding in the area with resultant repetition of the depositional sequence. The section is dominated by laterally continuous sandstone beds with minor mudstone and siltstone units.

Facies recognised at this location include B1.1, B2.1, C2.1,

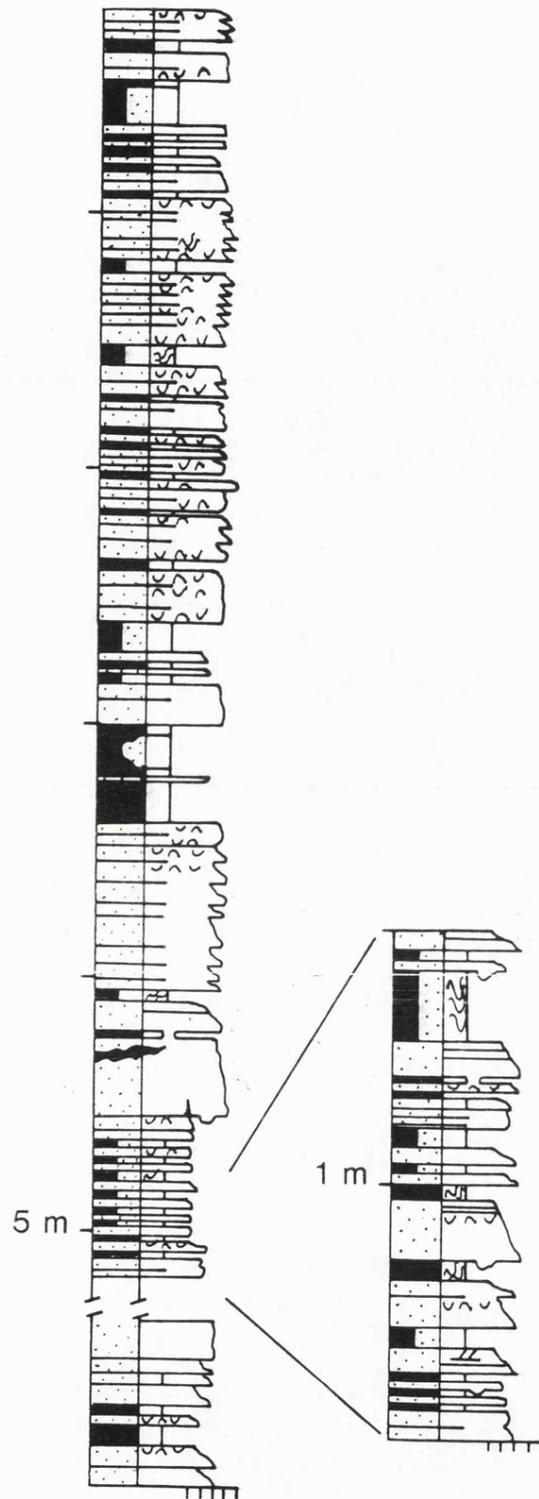


Figure 3.4 Measured section of Newport Sands Formation, Newport Sands.

C2.2, C2.3, (D2.1), (D2.3), (E1.1) and (E2.2).

Sandstone beds commonly show sharp bases and tops and are laterally continuous over tens of metres. Internally, they are structureless, graded and parallel laminated with typical Tabce, Tabe and Tbce Bouma (1962) sequences. Reverse grading may be present at the base of some beds. Amalgamation of beds is common as are dewatering structures, flame structures and clastic dykes. Many of the sandstone beds contain abundant mudstone rip-up clasts.

Siltstone beds are laterally continuous or discontinuous on a scale of tens of metres. Internally, they are normally graded and exhibit Tde Bouma (1962) divisions.

Mudstones are either structureless or parallel laminated.

### **3.3.3 Ceibwr Formation - Type locality: Carreg Yspar (SN 10224510)**

The Ceibwr Formation outcrops along the coast from just south of Careg Yspar (SN 100450) to Ceibwr Bay (SN 109458). It consists essentially of structureless and parallel-laminated mudstones with rare fine- to coarse-grained muddy sandstones. Both upper and lower contacts are gradational. Because of structural repetition and the essentially monolithologic character, the original thickness of the formation is difficult to determine.

The mudstones are commonly dark grey in colour with rare lighter coloured bands. At Ceibwr Bay some lenticular concretionary mudstone bands are found. The concretions are pyrite rich and are up to 2.0 x 20.0 cms. The concretionary layers are parallel to bedding.

The muddy sandstone layers are medium- to coarse-grained and laterally continuous over 12.0 m. The sandstone beds may be amalgamated. There is little evidence of any internal structure (see below).



### Ceibwr Formation - Section 1 (SN 10824585)

The section is situated on the northern limb of an anticline close to the top of the Formation at the gradational contact with the overlying Poppit Sands Formation (Fig. 3.5). The measured thickness is 3.8 m out of a total section thickness of 8.0 m.

Facies present include C1.1, D2.2, D2.3, E1.1, E2.1, and E2.2.

The sandstone beds are mud rich with abundant mudstone rip-up clasts, locally showing a crude parallel bedding. Bed bases range from parallel to erosional. Amalgamation is common. There is evidence of rippling on the tops of some of the beds.

Mudstones are either structureless or parallel laminated. The tops of individual mudstone beds tend to be oxidised and, therefore, lighter in colour.

Siltstone beds vary from laterally continuous to discontinuous over distances of 1.0 - 2.0 m. Internally, they display Bouma (1962) Tde lamination.

#### 3.3.4 Poppit Sands Formation - Type locality: Poppit Sands (SN 14604960)

The Poppit Sands Formation consists of intercalated fine- to coarse-grained sandstones, siltstones and mudstones with rare pebbly mudstones. The formation is c. 750 m thick and outcrops along the coast from Ceibwr Bay to Rees' Quarry (SN 170482) south of Gwbert. The base with the underlying Ceibwr Formation is coincident with the increase in the sandstone/mudstone ratio and is best seen at Ceibwr Bay (SN 109458) while the top is not exposed. The formation contains ichnofossils (Chondrites ichnosp., Helminthopsis ichnosp., Palaeophycus ichnosp. and Planolites ichnosp.), body fossils (pelmatozoan debris, trilobite and brachiopod fragments) the

graptolites Dicellograptus cf. forchammeri Geinitz, ?Dicranograptus sp. and Glyptograptus sp. (Collection site - SN 17524594). Based on the occurrence of Dicellograptus forchammeri Geinitz the formation can be assigned to the Dicranograptus clingani Zone of the Caradoc (Dr R.B. Rickards pers. comm.). Three sections (Figs. 3.6, 3.7, 3.8) have been logged in detail and these will be discussed below.

Poppit Sands Formation - Type locality: Section 1a,b (SN 14604960)

A coastal section approximately 70.0 m thick and situated along the beach from the harbour wall. The area is dominated by a large anticline and the two sections, 1a (Fig. 3.6) and 1b (Fig. 3.7), were logged on both limbs of the anticline, sharing a common basal bed, and are thus broadly coeval. The sections are separated by a lateral distance of 10.0 m although this may be exaggerated up-section. The two sections are dominated by coarse-grained sandstones and intercalated siltstones, mudstone and pebbly mudstone. The only recognised ichnofossil is Palaeophycus ichnosp.

Sedimentary facies types recognised at this locality are A1.3, A1.4, B1.1, B2.1, C2.1, C2.2, C2.3, D2.2, and E1.1.

The mudstones are dark and light grey in colour and commonly structureless. They are intercalated both with thinly- to very thinly-laminated siltstones, which are commonly sharp based and laterally continuous or discontinuous, and with fine-grained sandstones.

The fine-grained sandstones are sharp based, with minor loading, and some have rippled tops. Bouma (1962) T<sub>bce</sub>, T<sub>ce</sub> and T<sub>de</sub> are present. In Section 1a the pebbly mudstone facies (A1.3) is very common.

Sandstone beds are commonly medium- to coarse-grained with some granule grade sands being locally dominant. Internally the beds are structureless, parallel laminated, T<sub>ae</sub>, T<sub>abe</sub> and

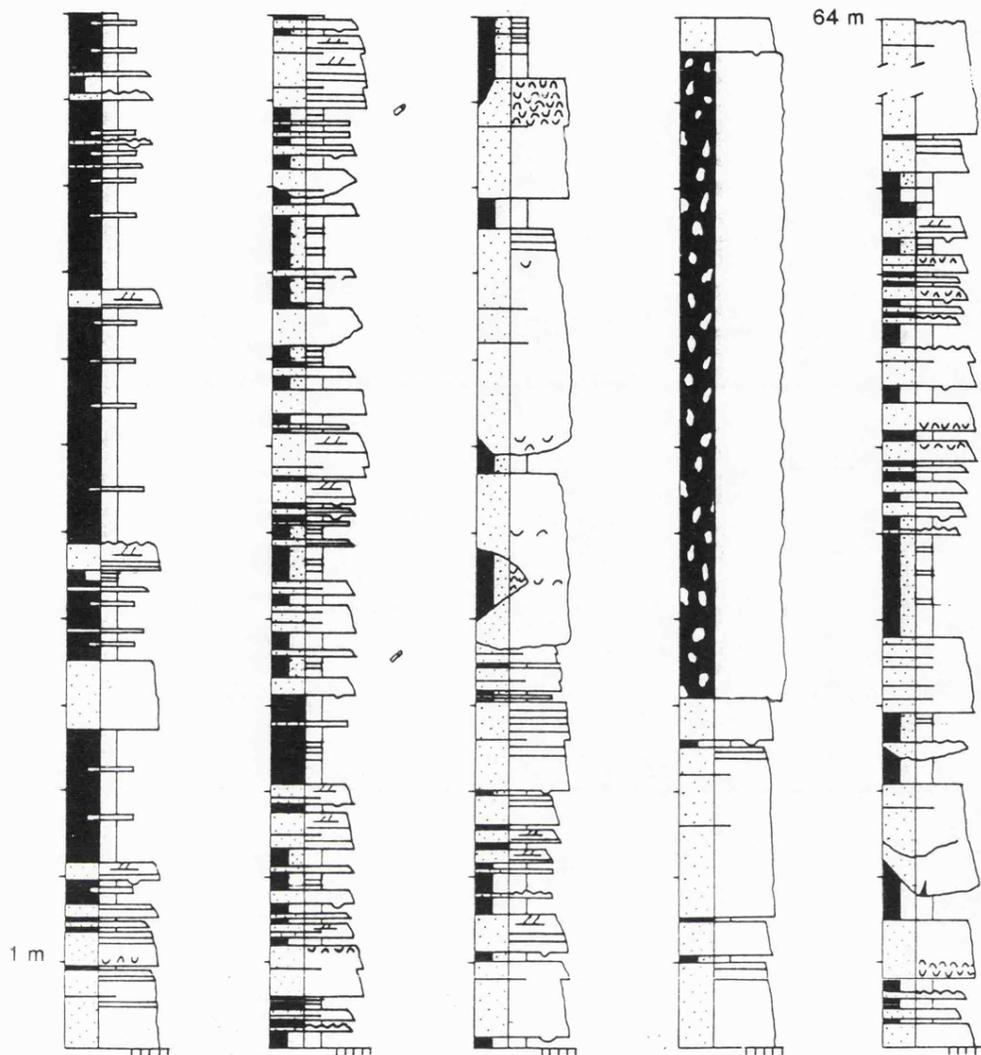


Figure 3.6 Measured section of Poppit Sands Formation, Poppit Sands (Section 1a - westerly section).

### 3.10

The. Mudstone rip-up clasts, where present, are concentrated towards the base or top or evenly distributed throughout the bed. Bed bases are erosive, sharp and amalgamated and commonly with flutes or load casts.

Correlations of sections show the erosive nature of some of the thicker sandstone units with channels cutting down through the underlying sediment by up to 5.0 m (Fig 3.8). Many of these thicker sandstones are seen to thin laterally and correlate with a series of thinner and finer-grained sandstones and siltstones.

#### Poppit Sands Formation - Section 2 (SN 17004818)

This section is located at Rees Quarry, near the top of the formation, close to the boundary with the overlying Gwbert Formation (Fig 3.9). The locality is a disused quarry exposing a thickness of 20.0 m, the top 3.25 m of which was logged in detail. The section is dominated by thin beds of fine- to medium-grained sandstone. Chondrites ichnospp. is the only ichnofossil preserved.

The sedimentary facies noted in this section include B2.1, C2.1, C2.2, C2.3, D2.1, D2.2 and E1.1.

The mudstones are commonly structureless and dark and light grey in colour. Intercalated siltstones are sharp based, fining upwards and laterally continuous to discontinuous. Individually, the beds are up to 2.0 cm thick.

Sandstones are fine- to medium-grained and normally graded. Internally they comprise Tbcde, Tcde, Tce and Tde divisions. Bases are sharp, with rare load and flute casts, while tops are rippled. Some of the rippled tops may be covered with mud drapes.

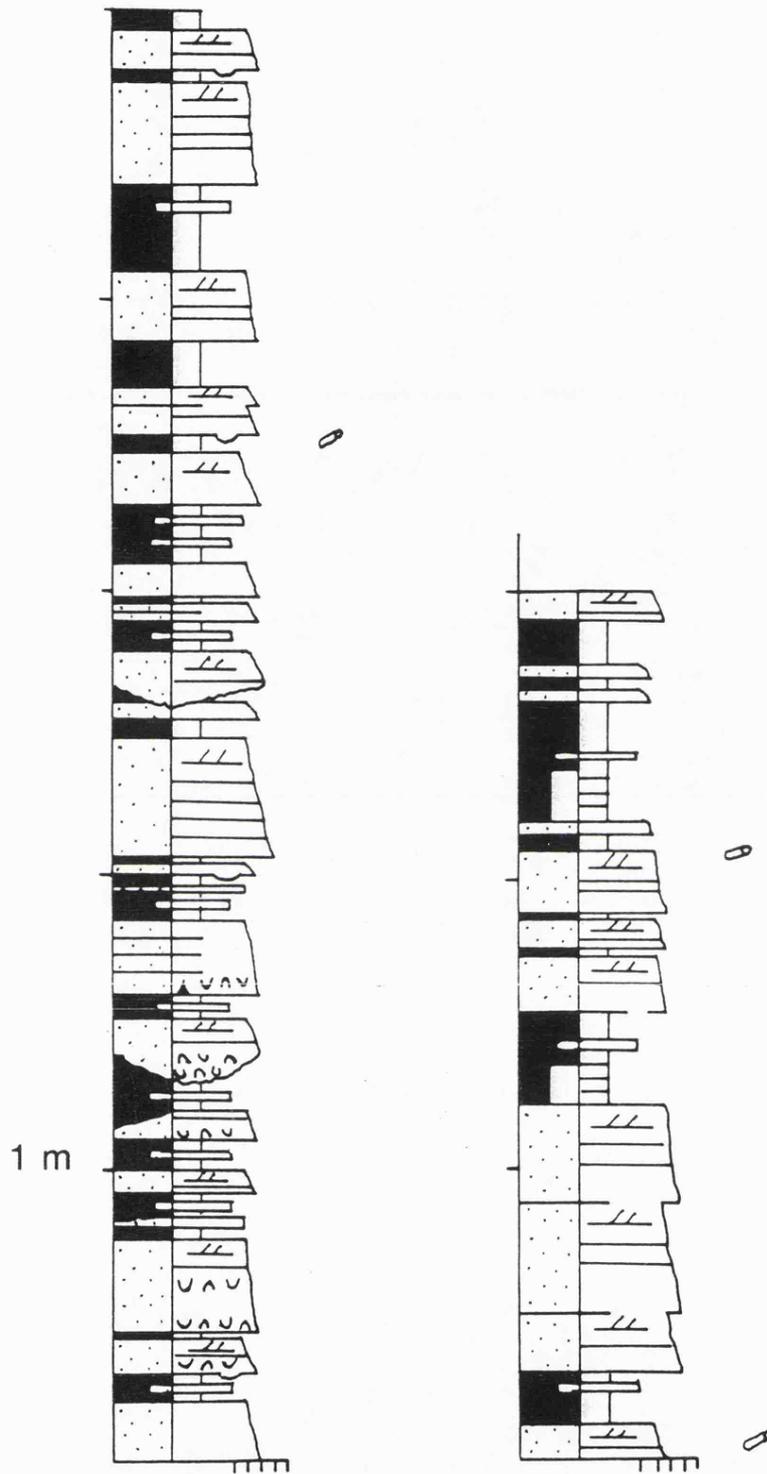


Figure 3.7 Measured section of Poppit Sands Formation, Poppit Sands (Section 1b - easterly section).

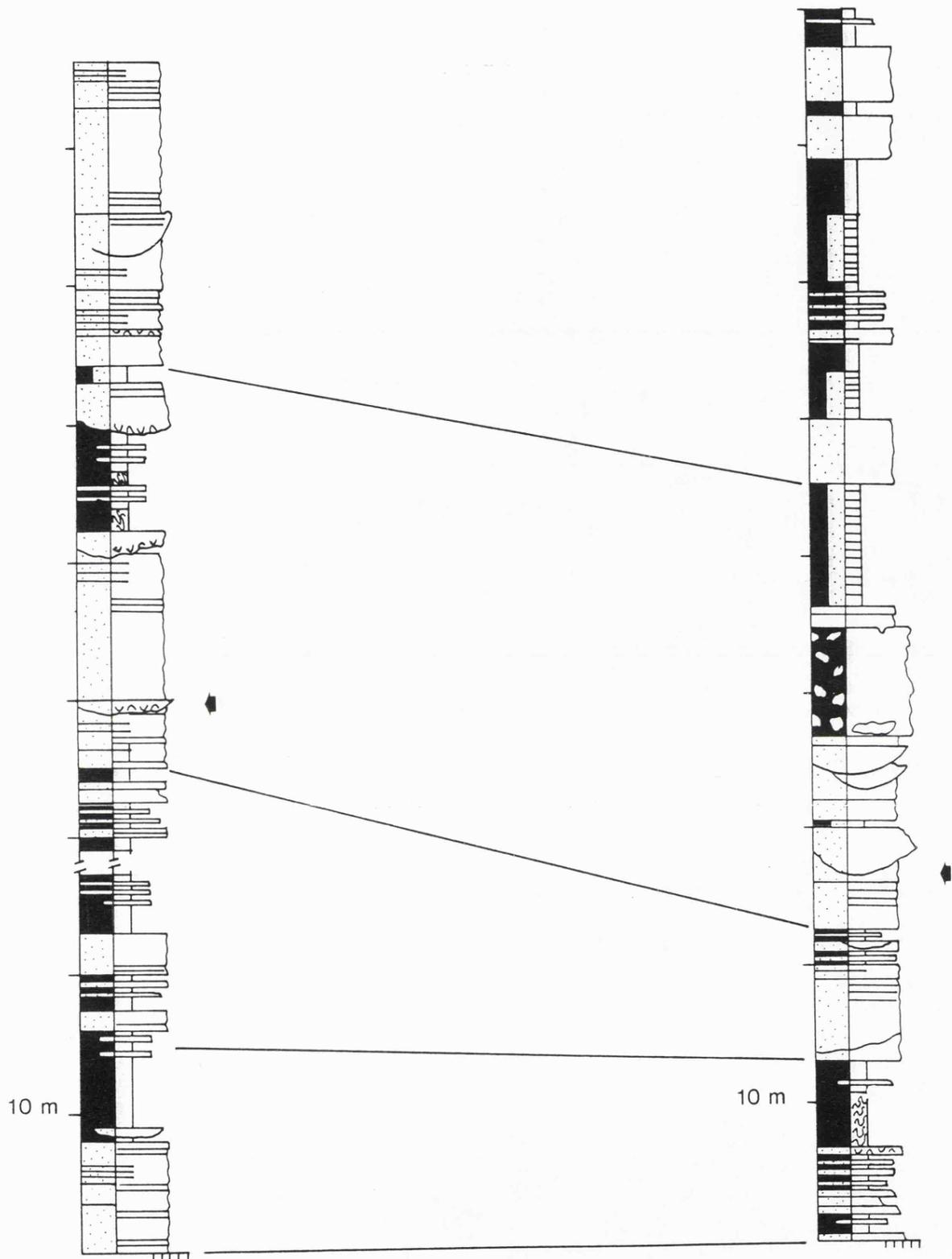


Figure 3.8 Measured section of Poppit Sands Formation, Poppit Sands correlating sections 1a and 1b. The arrows denote the position of the marker bioclastic beds (Facies Class A1.4).

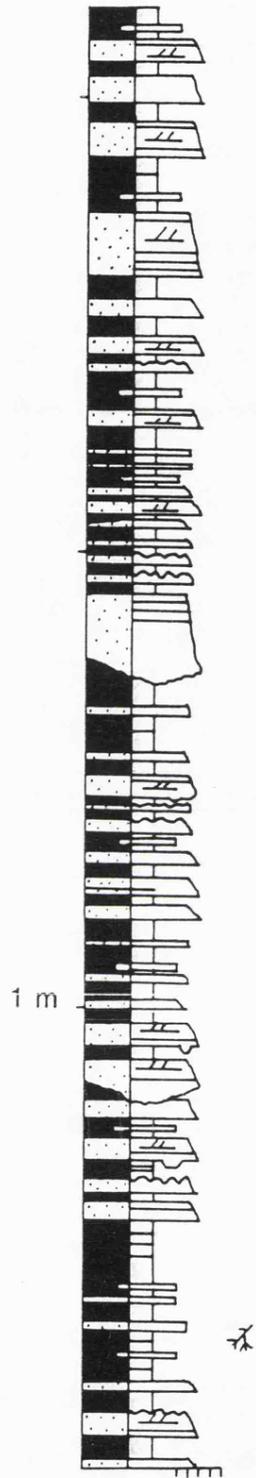


Figure 3.9 Measured section of Poppit Sands Formation, Rees Quarry.

## Part Two

## 3.4 Introduction

The dominant lithologies, areal extent and age (where known) of each of the formations outcropping between Cardigan and New Quay are briefly outlined. Representative sections are then described and interpreted for each of the formations.

## 3.4.1 Gwbert Formation

The formation comprises dark and medium grey banded mudstones with rare, intercalated fine- to medium-grained sandstone and siltstone beds (Craig 1985). No graptolites have been reported from the formation. The formation crops out between Craig-y-Gwbert (SN 158502) and Traeth Mwnt (SN 194519).

## Gwbert Formation - Section 1 (SN 15835014)

The section is located on the coast immediately to the south of the hotel and is approximately 2.5 m thick, 0.5 m of which was logged in detail (Fig. 3.10). The sequence is dominated by structureless and parallel-laminated mudstones with minor siltstone bands. Two ichnogenera are present - Chondrites ichnospp. and Planolites ichnosp.

A variety of facies types are recognised at this locality, including C2.3, D2.1, D2.2, D2.3, E1.1 and E2.1.

The mudstones are lithologically similar and comprise two types differentiated by colour. The dark grey mudstone, representing the unoxidised mudstone, is always positioned above the siltstone and sandstone beds while the lighter coloured mudstone occurs elsewhere. The boundaries between the two mudstone types tend to be diffuse.

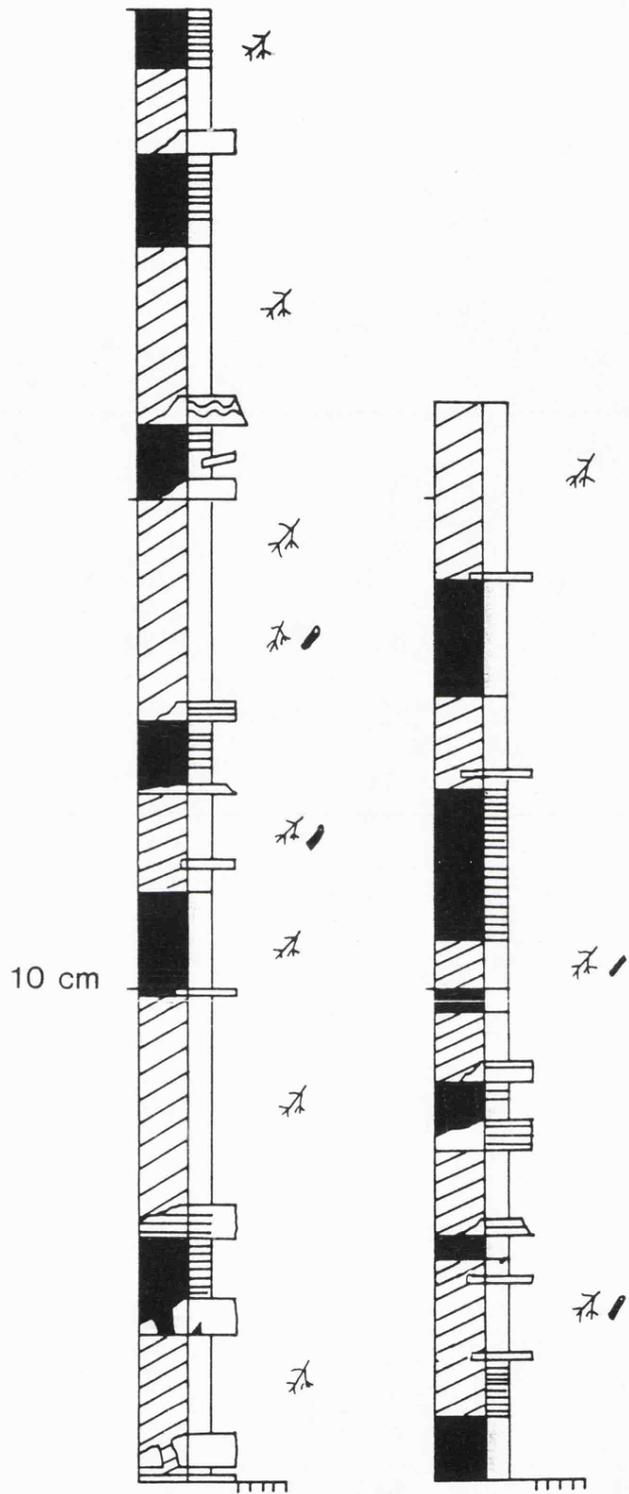


Figure 3.10 Measured section of Gwbert Formation, Gwbert (Section 1).

### 3.12

Siltstones are light grey with sharp bases which are rarely gradational and undulating. Tops commonly grade upwards into parallel laminae which may be convolute and draped with dark grey mudstone. Bottom structures include ball and pillow structures and associated flames.

Sandstones are fine-grained and laterally discontinuous. Internally they grade upwards with asymmetrically cross-laminated tops.

#### Gwbert Formation - Section 2 (SN 16185121)

This section is stratigraphically higher than the previous section and is positioned in a small cove adjacent to the golf course (Fig. 3.11). It is approximately 3.0 m thick of which the basal 0.9 m was logged in detail. Ichnofossils are relatively abundant, particularly in the mudstones, and both Chondrites ichnospp. and Planolites ichnospp. were recognised.

Facies types recognised at this locality include C2.3, D2.1, D2.2, D2.3, E1.1 and E2.1.

The section is dominated by mudstones which are both structureless and parallel laminated, although coarser-grained facies are more prevalent than in Section 1 of the Gwbert Formation. Either dark- or medium-grey mudstones are present. The dark-grey mudstone commonly contain very thinly-laminated siltstone bands. The medium-grey mudstone is either structureless or contains thin- to very thin-laminated siltstone bands which are parallel and sub-parallel to bedding. Profiles which grade upwards from zones of intense bioturbation to zones with no apparent evidence of bioturbation are generally found within the medium-grey mudstones. Basal contacts are sharp while upper contacts are diffuse. The medium-grey mudstone may grade upwards into dark-grey mudstone.

Siltstone bands are commonly lenticular. They are either continuous or discontinuous on a scale of metres. Tops are



graded while bases are sharp. A variety of bottom structures are present including loading, ball-and-pillow structures, and flames. Internally, the siltstones commonly display Tbce and Tce divisions. Some laminae may show evidence of folding as a result of wet-sediment deformation.

The sandstones are fine-grained with sharp bases which may be loaded. Tops are graded and internally the sandstones may be classified according to Bouma (1962) as Tabe and Tbce. The Tc divisions may be convolute.

#### 3.4.2 Mwnt Formation

The Formation comprises interbedded fine-grained sandstones, siltstones and mudstones. The dominant lithology of banded mudstones with thin siltstone laminae is broadly similar to both the underlying Gwbert and overlying Tresaith formations which crop out on either side of the Mwnt Formation. The Mwnt Formation crops out along the coastline from Traeth Mwnt (SN 194519) to south of Cribach Bay (SN 252523).

The mudstones are dark and medium grey in colour. The darker mudstones are associated with the fine-grained sandstones and siltstones and range from parallel-laminated to structureless. The lighter coloured mudstones may contain isolated burrows of Planolites ichnosp. and burrow systems of Chondrites ichnosp. They are internally structureless or parallel laminated.

Siltstones are lenticular and vary from continuous to discontinuous over lateral distances of 10's of metres. They are sharp based with graded tops which grade upwards into lenticular siltstone beds intercalated with dark grey mudstone. This, in turn, grades upwards into medium grey mudstone. Bottom structures include load casts and ball-and-pillow structures.

Craig (1985) has reported the occurrence of medium-grained scoured sandstones which are up to 15.0 cm in depth. The

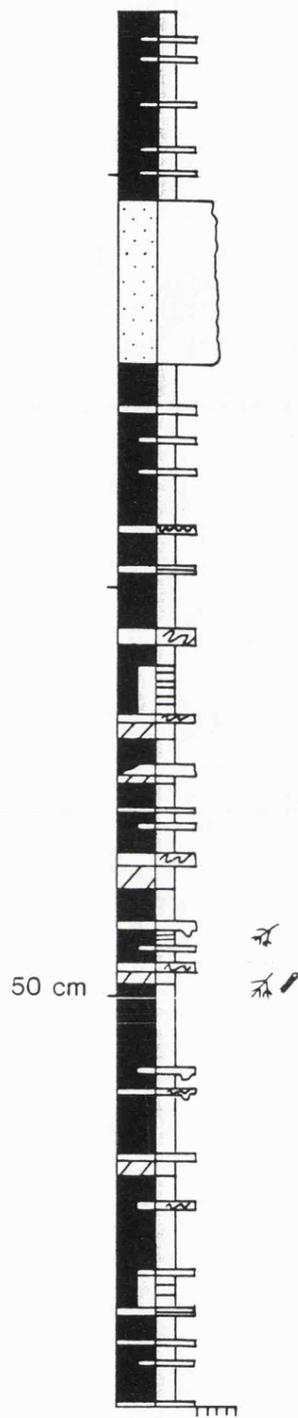


Figure 3.12 Measured section of Mwnt Formation, Mwnt.

scour infill is parallel laminated and contorted.

#### Mwnt Formation - Section 1 (SN 19325195)

A coastal section, approximately 14.0 m thick, of which 1.7 m was measured in detail (Fig 3.12). The ichnofossils Chondrites ichnospp. and Planolites ichnosp. are recorded.

Facies present include C2.2, C2.3, D2.1, D2.2, E1.1 and E2.1.

Sandstones are coarse- to fine-grained and commonly exhibit Bouma (1962) Tabe, Tbce and Tce divisions. The Tc divisions may be convolute.

Mudstones are parallel laminated, graded and have both oxic and anoxic layers.

Siltstone beds are laterally continuous or discontinuous over lateral distances of tens of metres. Internally the beds are Tab, Tabc and Tde. The Tc-division may be convolute. Sliding, slumping, loading and wet-sediment deformation are common. Internally, some beds pass laterally from being structureless to parallel laminated.

#### 3.4.3 Tresaith Formation

The Tresaith Formation outcrops from just south of Cribach Bay (SN 252523) to the northern end of Penbryn Beach (SN 288522). It is broadly similar to the two underlying formations with banded mudstones and minor intercalated siltstone and fine-grained sandstone laminae as the dominant lithologies. Bioturbation is locally abundant with Chondrites ichnospp. and Planolites ichnosp. recorded. As with the underlying formations, occurrence of these ichnofossils is restricted to the lighter coloured mudstone bands. A graptolite fauna of Dicellograptus anceps Zone has been reported (Anketell 1987) thus placing the formation firmly in the Ashgill.

The mudstones are either dark or medium grey in colour. They are identical to similar banded mudstones in the Mwnt and Gwbert formations. The mudstones at the northern end of the formation, at Penbryn Beach, contain mudstone concretions which are oriented either parallel or sub-parallel to bedding. Concretions with outer skins of cone-in-cone structures are also present.

Both the siltstone and sandstone laminae are thin bedded to very thin bedded and are either laterally continuous or discontinuous over distances of tens of metres. The maximum observed thickness is 1.8 cm although they are commonly less than 0.35 cm. Beds are sharp based with graded tops. They are internally parallel or cross laminated.

Three sections were measured. Two of these are located near the base of the formation while the third is located near the top.

#### Tresaith Formation - Section 1 (SN 25885151)

The section is located in the harbour of Aberporth and is approximately 15.0 m thick, of which 1.1 m was measured in detail (Fig 3.13). The ichnofossils Chondrites ichnospp. and Planolites ichnosp. are very common.

Facies recognised include D2.2, E1.1 and E2.1.

Mudstones are graded, structureless or parallel laminated. The tops of the mudstone beds are lighter coloured, suggesting deposition under oxic conditions. Bases tend to be sharp and may be silty with rare small washouts noted. Early diagenetic concretions, arranged parallel to bedding, are present.

Siltstones are commonly sharp based with gradational tops. Bottom structures, for example flames, load casts and ball-and-pillow structures, are recorded. Internally, the siltstones are structureless or display Tde units.

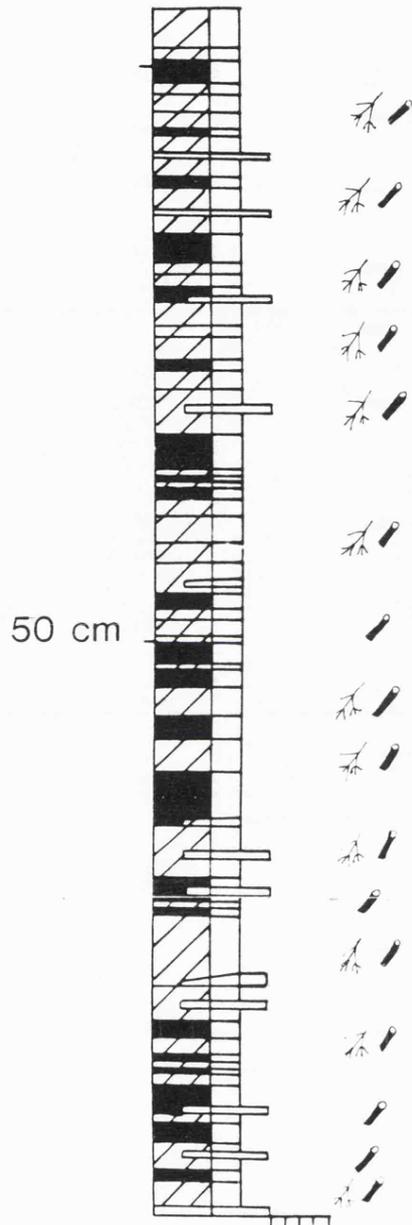


Figure 3.13 Measured section of Tresaith Formation, Aberporth (Section 1).

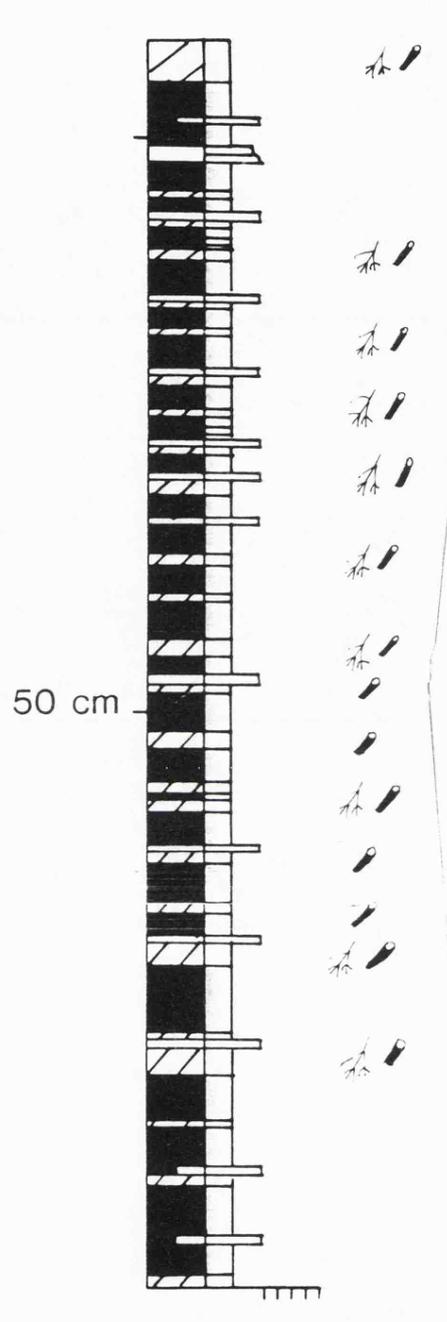


Figure 3.14 Measured section of Tresaith Formation, Aberporth (Section 2).

**Tresaith Formation - Section 2 (SN 25735165)**

This coastal section stratigraphically overlies the previous section (Fig 3.14). Measured thickness is 1.2 m.

Facies present include D2.1, D2.2, E1.1 and E2.1.

Mudstones are similar to those of Section 1. Siltstone beds, however, are more common. They are sharp-based with graded tops. The siltstone beds grade laterally from being either normally graded or structureless to exhibiting parallel lamination (Tde).

**Tresaith Formation - Section 3 (SN 27605154)**

This section is situated near the top of the Tresaith Formation at Tresaith (Fig 3.15). It is a coastal section 4.0 m thick of which 1.4 m was measured in detail.

Facies present include (D2.1), (D2.2), E2.1 and E2.2.

Siltstone beds are rare. They are sharp based and grade upwards. They commonly grade laterally from being structureless to parallel lamination (Tde).

Mudstone beds dominate the section. There are two main types: the graded muds commonly have a silty base which, in turn, may be parallel laminated or graded. The laminated mudstones are unbioturbated and have a high organic carbon content (see Chapter Six).

**3.4.4 Llangranog Formation**

The Llangranog Formation comprises mudstones, siltstones and sandstones. It outcrops from north of Penbryn Beach (SN 288522) to north of Llangranog village (SN 310542). The Formation has been subdivided into five members which individually show rapid facies and thickness variations

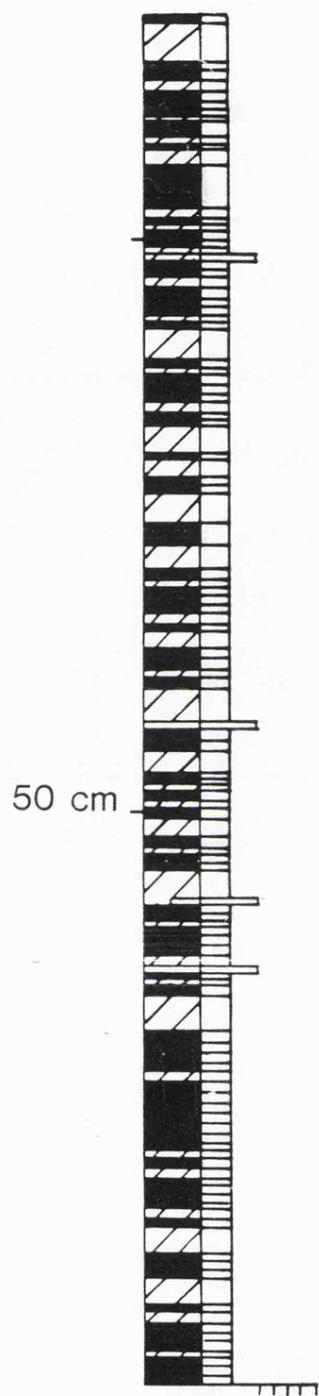


Figure 3.15 Measured section of Tresaith Formation, Tresaith (Section 3).

(Anketell 1987). Based on graptolite data, the formation may be dated as Ashgill (Hendricks 1926; Anketell 1963). Ichnofossils are common, with the following ichnogenera reported Chondrites ichnospp., Circulichnis ichnosp., Cochlichnus ichnosp., Gordia ichnosp., Helminthopsis ichnosp., Palaeophycus ichnosp., Planolites ichnosp. and Protopaleodictyon ichnosp.

The lowermost Penbryn Beds comprise interbedded sandstone and mudstone and grade upwards into the Carreg-y-Ty Beds which are mostly mudstone with minor sandstone lenses. Wet-sediment deformation is common in this member (Craig 1985). The overlying Sarnau Beds are poorly developed on the coast but have been mapped inland by Anketell (1963) and Hasso (1974). They comprise mudstones, siltstones and rare lenses of fine-grained sandstone. The member is lithologically similar to the overlying Morfa Beds which contain coarser sandstone fractions.

The most widespread and stratigraphically the highest member in the formation is the Traeth Bach Beds. These are well exposed in the vicinity of Llangranog village and comprise banded mudstones and intercalated fine- to medium-grained sandstones (see below). Wet-sediment deformation, consisting of numerous slide and slump sheets, is well developed towards the base of the member. Analysis of the vergence directions of the slide folds by Craig (1985) and Anketell (1963) confirms their common direction, which is to the north west. The intensity of deformation decreases up section. The sandstone:mudstone ratio also changes up section with sandstone beds becoming finer grained and thinner bedded. The deformed sediment areas are intercalated with zones of little or no wet-sediment deformation.

#### Llangranog Formation - Section 1 (SN 31095426)

The dominant feature of this section is the chaotic nature of the sediment deformation (Fig 3.16). Folds are commonly truncated by overlying beds or their wavelength may decrease



up section. Many of the folds are rootless. Other features consistent with a wet-sediment origin, for example ball-and-pillow structures and syn-sedimentary faults, are common. The section is situated on the coast and 80.0 m was measured of a total section thickness of 110.0 m.

Facies recognised include B2.1, C2.1, C2.2, C2.3, D2.1, (E2.1), (E2.2), F2.1 and F2.2.

Sandstone beds are sharp based with graded tops. Internally, the beds are parallel laminated or exhibit Bouma (1962) Tbce and Tce divisions.

Siltstone beds are graded and stratified. Internally, they exhibit Tbce and Tce divisions.

Mudstones are rare but where present vary from structureless to parallel laminated.

#### 3.4.5 Gaerglwyd Formation

The Formation comprises dark and medium grey banded mudstones with intercalated siltstones and rarer fine-grained sandstone beds. A number of graptolite assemblages have been reported from the formation (Anketell 1987). The base of the formation contains Glyptograptus persculptus, Climacograptus scalaris var. normalis and C. scalaris var. miserabilis. The presence of G. persculptus dates the formation at the base of the Silurian (Lower Llandovery). Overlying assemblages (within the Gaerglwyd Formation) belong to the Akidograptus acuminatus Zone and the base of the Monograptus triangulatus Zone (Anketell 1963).

The formation outcrops along the coastline from Ynys Loctyn (SN 316551) to the north of Traeth Gaerglwyd (SN 332558). Septarian nodules have been collected from the base of the Formation (Craig 1985). Chondrites ichnospp. is the only recorded ichnofossil.

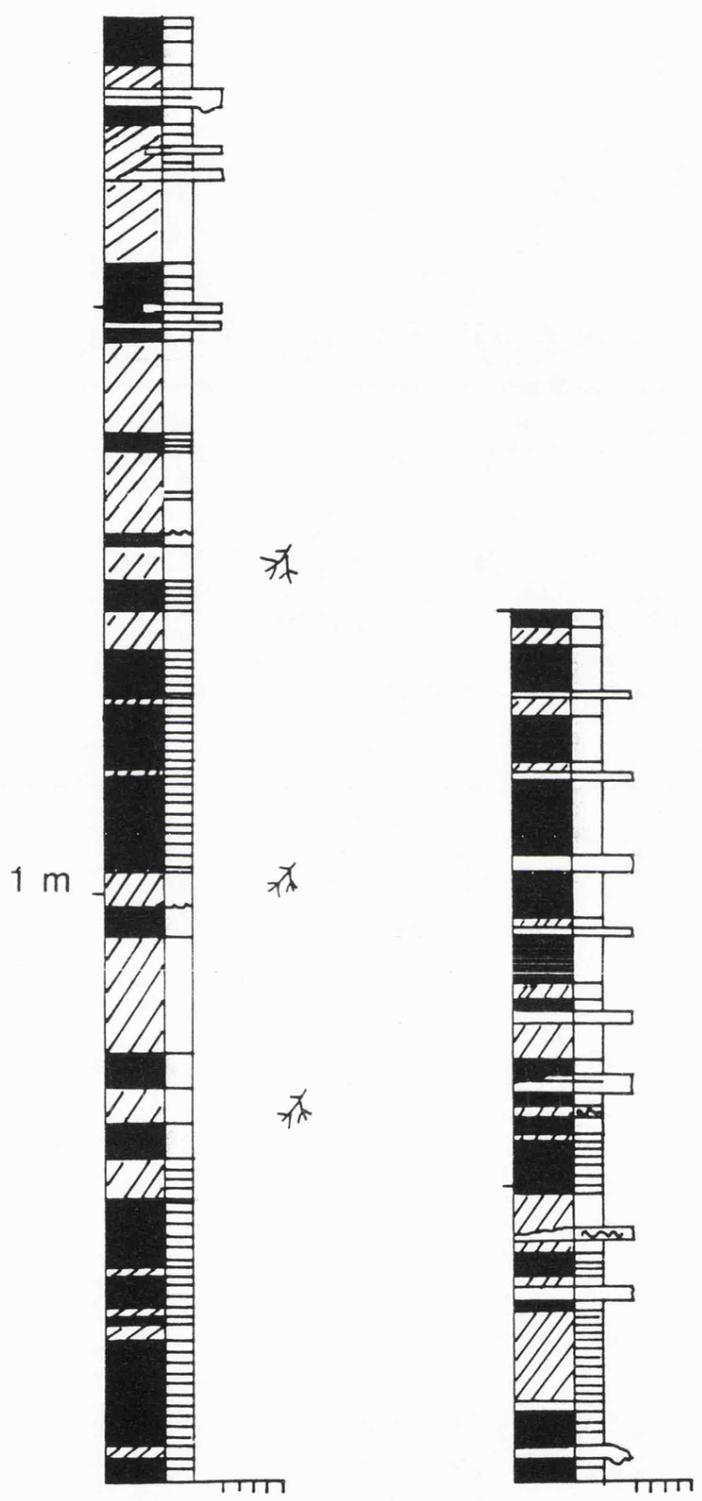


Figure 3.17 Measured section of Gaerlwyd Formation, Traeth-yr-Ynys.

### Gaerglwyd Formation - Section 1 (SN 31905495)

The section is approximately 40.0 m thick, of which 4.0 m was logged in detail (Fig 3.17). The sequence is dominated by structureless and parallel laminated mudstones with thin to very thin bedded siltstones becoming increasingly important towards the top to the section.

The facies type recognised at this locality include D2.1, D2.2, D2.3, E1.1, E2.1 and E2.2.

The mudstones are either dark or medium grey in colour and are internally structureless or parallel laminated. The darker mudstone is sharp based with a graded top which may grade into lighter coloured mudstone. The medium grey mudstones contain disseminated pyrite and local evidence of bioturbation. Bed bases are sharp or gradational.

Siltstones are lenticular and continuous to discontinuous over lateral distances of 10's of metres. Internally they are structureless and parallel bedded. Bases are sharp but may be loaded with internally laminated pseudonodules of siltstone being produced.

#### 3.4.6 Allt Goch Formation

The Allt Goch Formation consists of well bedded, fine- to medium-grained sandstones interbedded with minor siltstone and mudstone. Anketell (1963) noted that the sandstone:mudstone ratio decreased inland. The Formation outcrops along a ridge stretching from Ynys Loctyn (SN 316551) to Allt Goch (SN 319542).

Both Hendricks (1926) and Anketell (1963) have reported graptolite assemblages from the interbedded mudstones. On the basis of these the formation may be placed in the Monograptus triangulatus Zone. The presence of a shelly fauna, dominated by Skenidioides sp., in some coarser sandstone washouts confirms the lower Llandovery date.

Graptolites collected in the present study, including Dictyonema sp., Monograptus austens Hutt, M. austens Hutt (?sequens or ?praecursor), M. t. triangulatus (Harkness), Petalograptus cratoelongatus (Kurck), Rastrites longispinus (Perner) and Rhaphidograptus toernquisti (Elles & Wood), may also be assigned to the triangulatus Zone (Dr R.B. Rickards pers. comm.; Collection site - SN 31815411). Recorded ichnofossils include Chondrites ichnospp., Helminthopsis ichnospp., Palaeophycus ichnospp. and Planolites ichnospp.

The sandstones are sharp based and may be classified as Bouma (1962) Tbcde, Tcde, and Tde, with rare Ta, divisions. Tc divisions are commonly convolute. Bottom structures include load, flute and groove casts. The interbedded mudstones and siltstones are commonly parallel laminated and the siltstones may show poorly developed Tbce and Tce units.

Allt Goch Formation - Section 1 (SN 31785419)

The section is located at the side of a hill overlooking the church at Llangranog (Fig 3.18). The total exposed thickness is 18.0 m which is logged in detail.

The following facies are recorded at this locality; (C2.1), C2.2, C2.3, D2.1, D2.2, E1.1, (E2.1), and (E2.2).

Sandstone beds are sharp based although small washout structures are present. The base of some sandstone beds comprise cone-in-cone structures. The tops of the beds are either sharp or gradational. Internally they display all of the Bouma (1962) divisions although the basal Ta division is rare and the Tc division is typically convolute. Synsedimentary faulting is common as is bed amalgamation.

Siltstone beds are sharp based and grade upwards. Internally, they comprise Tbce and Tce divisions, and, as in the sandstones, the Tc divisions are typically convolute.

Mudstones are generally laminated, graded or structureless.

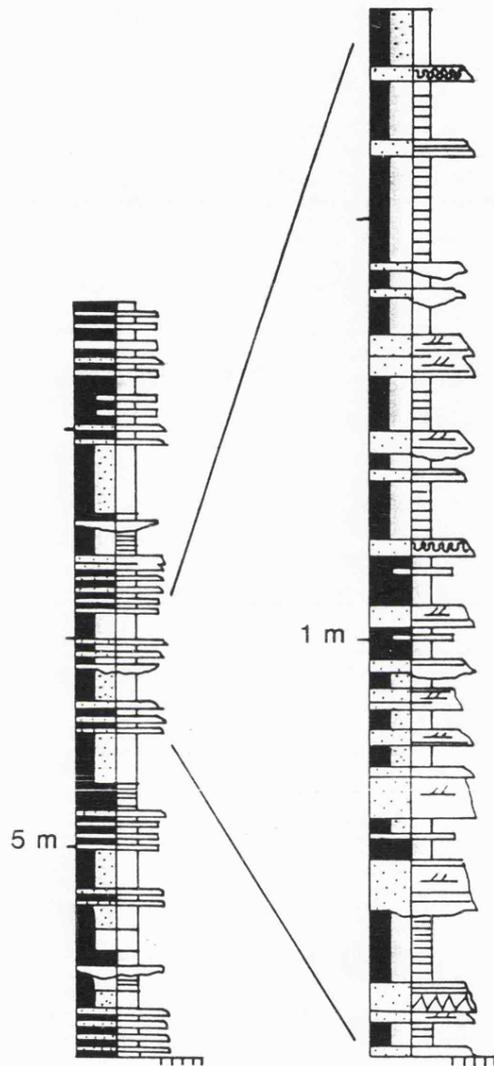


Figure 3.18 Measured section of Allt Goch Formation,  
Llangranog.

Orthoconic nautiloids are common.

#### 3.4.7 Cefn Cwrt Formation

The formation comprises banded mudstone with minor intercalated siltstone and rare fine-grained sandstone. It outcrops along the coast from Traeth-y-Gaerlwyd (SN 332558) to Cefn Cwrt (SN 328553). The sandstones may be cross stratified and convolute (Craig 1985). Graptolitic data suggests that much of the formation lies within the upper Llandovery Monograptus convolutus Zone (Hendricks 1926; Anketell 1963). Chondrites ichnospp. and Planolites ichnosp. are the only recorded ichnofossils. The formation was not examined in detail in this study.

#### 3.4.8 Llangraig Formation

The formation consists of banded mudstones and intercalated siltstones with rare fine-grained sandstones. It is exposed in the cliffs around Trwyn Croi (SN 332557). Concretions with skins of cone-in-cone structures have been reported (Craig 1985). Anketell (1963) collected four graptolite species which place the formation in the upper Llandovery Monograptus sedgwickii Zone. Chondrites ichnospp. and Planolites ichnosp. are the only recorded ichnofossils. The formation has not been examined in detail in this study.

#### 3.4.9 Grogal Formation

The Grogal Formation comprises banded mudstone, siltstone and sandstone and is well exposed in the area of Cwm Tudu (SN 355576). Poorly preserved graptolites were reported by Anketell (1963) but these were not identifiable. Chondrites ichnospp. and Planolites ichnosp. are the only recorded ichnofossils.

The mudstones range from light to dark grey in colour and

contain disseminated pyrite. Thin- to very thin-bedded bedded siltstones are both laterally continuous and discontinuous. Bases are sharp while tops are graded and parallel or cross laminated. The fine- to medium-grained sandstones are sharp based with some bottom structures such as loads and ball-and-pillow structures. Internally, the sandstones are parallel to cross laminated.

#### Grogal Formation - Section 1 (SN 35535758)

A coastal section approximately 20.0 m long of which 3.55 m was measured in detail located on the southern end of Cwm Tydi Bay (Fig 3.19). Facies recognised include (C2.2), (C2.3), D2.1, D2.2, E1.1 and E2.1.

Mudstones are dark grey in colour grading up into a lighter coloured mudstone. Parallel lamination is also noted. Bioturbation is rare.

Siltstone beds are sharp-based and laterally continuous to discontinuous over distances of metres. Internally, they display Tcde and Tde divisions.

Sandstones are rare. The beds are sharp based and grade upwards from medium to finer grained. Bouma (1962) Tace, Tbce, Tbe, Tcde and Tce divisions are present while some beds are structureless. The tops of many of the sandstone and siltstone beds show some evidence of current reworking.

#### Grogal Formation - Section 2 (SN 37225934)

A coastal section at Craig Coybal of which 2.8 m was measured in detail (Fig. 3.20). Facies recorded at this locality include (C2.2), (C2.3), (D2.1), D2.2, E1.1, E2.1 and (E2.2).

Mudstones range from graded to parallel laminated. The top portion of many of the mudstone units is lighter in colour. Rare early diagenetic concretions are present and are

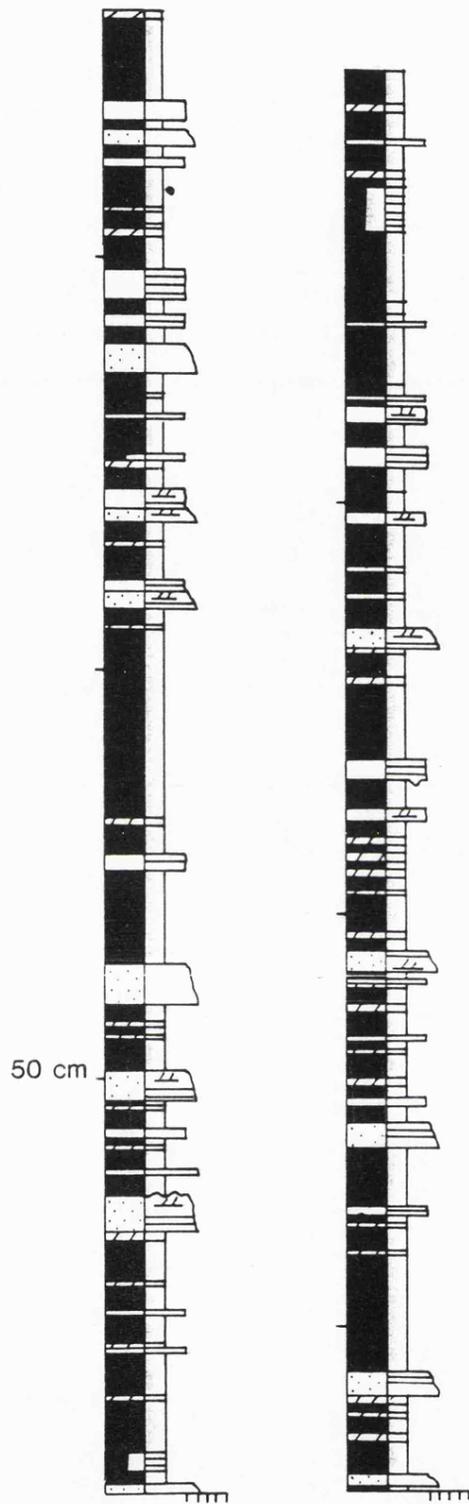


Figure 3.19 Measured section of Grogal Formation, Cwm Tudu (Section 1).

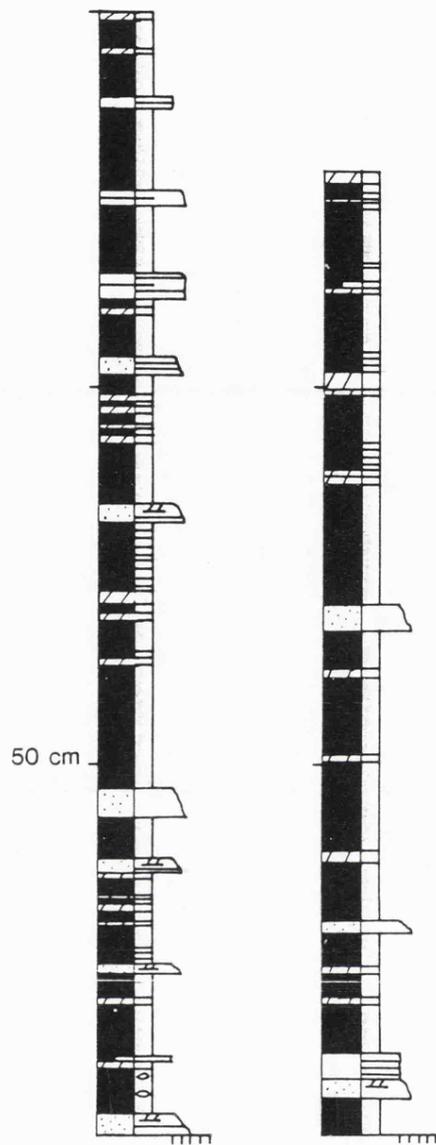


Figure 3.20 Measured section of Grogal Formation, Craig Coybal, (Section 2).

parallel to bedding.

Siltstones are rare, laterally discontinuous and sharp-based with graded tops. Internally, they may be classified as Tde.

Sandstones are sharp based with reworked tops. Internally, they contain Tbc and Tce Bouma (1962) divisions. The sandstones are very quartzose and have a 'clean' appearance.

#### 3.4.10 Aberystwyth Grits Formation

The Aberystwyth Grits Formation has been described in detail by Wood and Smith (1959) while a number of subsequent authors have focussed on more specific aspects such as the structure (Price 1962) and the ichnology (Crimes and Crossley 1980). The formation comprises interbedded sandstones, siltstones and mudstones and is up to 240-300 m thick in the Aberystwyth area (Cave & Hains 1986). It crops out in a crescentic belt from New Quay (SN 388605) in the south to Borth (SN 605888) in the north. Graptolite faunas of the formation belong to the upper Llandovery Monograptus turriculatus Zone (Wood & Smith 1959; Cave & Hains 1986). The formation contains the most diverse ichnofaunal assemblage in the study area, including, Chondrites ichnosp., Circulichnis ichnosp., Cochlichnus ichnosp., Cosmorhapha ichnosp., Desmograpton ichnosp., Gordia ichnosp., Helminthoida ichnosp., Helminthopsis ichnosp., Nereites ichnosp., Neonereites ichnosp., Palaeophycus ichnosp., Paleodictyon (Glenodictyon) ichnosp., Paleodictyon (Squamodictyon) ichnosp., Planolites ichnosp., Protopaleodictyon ichnosp. and Spirophycus ichnosp.

#### Aberystwyth Grits Formation - Section 1 (SN 38526048)

The section is located along the coast south of New Quay and is approximately 60.0 m thick of which the top 34.0 m was logged in detail (Fig. 3.21). The section is dominated by the coarse granule graded sandstones. One specimen of Palaeophycus ichnosp. was noted.

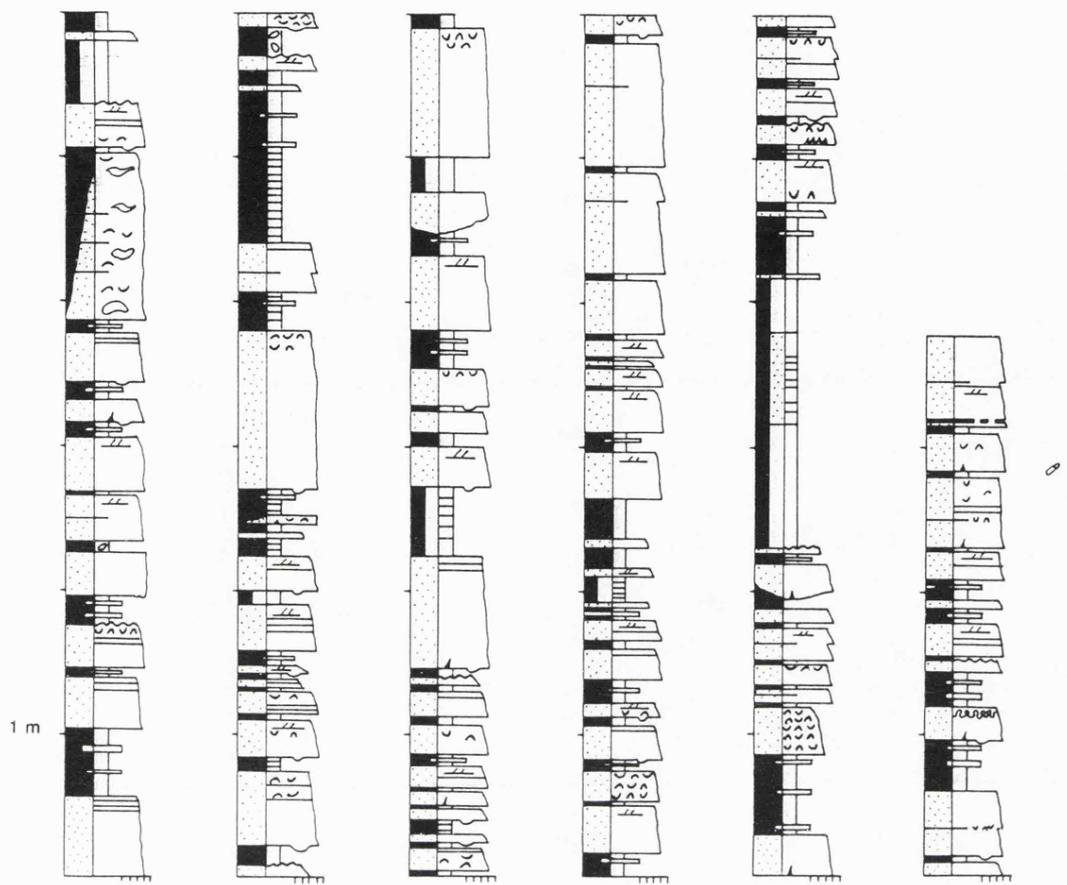


Figure 3.21 Measured section of Aberystwyth Grits Formation, New Quay (Section 1).

A number of facies types were recognised at this locality including B1.1, B2.1, C2.1, C2.2, C2.3, D2.1, D2.2, E1.1 and F2.2.

The mudstones are either structureless or parallel laminated. They may contain sandstone, siltstone or mudstone intraformational rip-up clasts up to 6.0 x 90.0 cm and oriented parallel or subparallel to bedding. Internally, these rafts are bedded and they may be folded or kinked. The larger clasts are located towards the tops of the beds while the smaller fragments are distributed throughout.

The siltstones are lenticular and laterally continuous to discontinuous over distances of tens of metres. Internally, many of the beds may be classified according to Bouma (1962) as Tabce, Tabe, Tbce, Tcde, Tce and Tde. The bases are sharp while the tops are rarely rippled.

Sandstones are fine- to granule-grade with granules, if present, being concentrated at the base of the bed as a lag, or close to the base at the top of a reverse graded basal layer. Bed bases are commonly sharp with a variety of bottom structures such as grooves, flutes, scours, flames, ball-and-pillow structures, sandstone dykes and loading. Internally, many of the sandstone beds are Tabce, Tabe, Tace, Tae and Tbe. The Tc divisions may be convolute. Some sandstone beds contain mudstone or siltstone intraformational clasts oriented parallel or subparallel to bedding. The clasts are located at the top, bottom or throughout the beds. Bed amalgamation is common.

#### Aberystwyth Grits Formation - Section 2 (SN 38686041)

A quarry section located just south of New Quay which is up to 30.0 m thick of which the top 11.0 m was logged in detail (Fig. 3.22). The section is dominated by coarse- to granule-grade, thick-bedded sandstones although there is evidence both of upward fining and thinning with a decrease in the sandstone:mudstone ratio up section. Palaeophycus

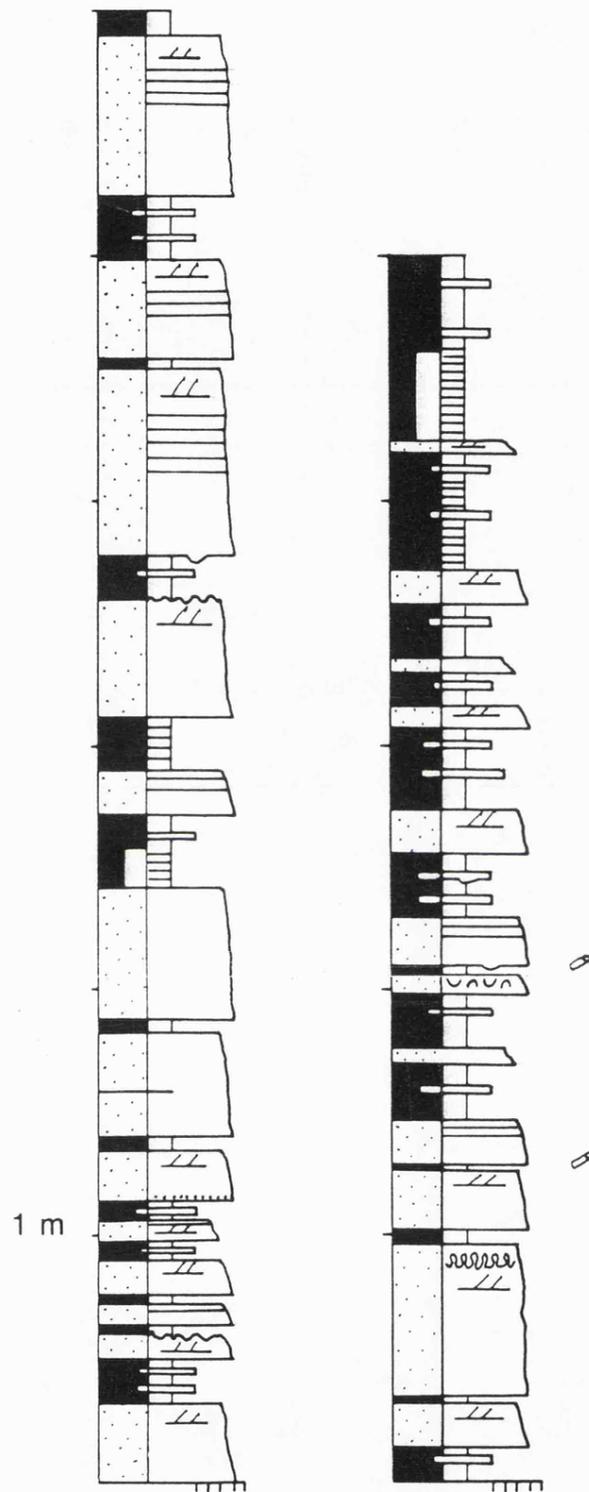


Figure 3.22 Measured section of Aberystwyth Grits Formation, New Quay Quarry (Section 2).

ichnosp. is the only ichnofossil.

The sedimentary facies reported from this locality include B1.1, B2.1, C2.1, C2.2, C2.3, D2.1, D2.2 and E1.1.

The mudstones range from structureless to parallel laminated. The intercalated siltstones are thin- to very-thinly laminated and are both lenticular continuous to discontinuous over distances of metres. Internally they are Tcde, Tce and Tde. Bases are commonly sharp although there is minor loading.

Sandstones are fine- to coarse-grained with granule grade fragments located as lags or at the tops of reverse grading sequences. Amalgamation of the sandstone beds is common. Internally, the sandstones display Tabce, Tabe, Tacde and Tace divisions, where the Tc division is rarely convolute. Bases are sharp with minor loading. Rafts of cross-laminated sandstone are distributed throughout one sandstone bed.

#### Aberystwyth Grits Formation - Section 3 (SN 39315957)

A sandstone-dominated section located just north of New Quay harbour, on the northern side of the anticline that underlies the village of New Quay, was measured (Fig. 3.23). The sediments in this area are the lateral equivalents of the coarser sediments in the previous two sections. The measured thickness of the section is 7.6 m. A variety of ichnofossils were noted including Gordia marina, Helminthopsis ichnosp., Palaeophycus tubularis, Palaeophycus ichnosp. and Paleodictyon (Glenodictyon) strozzi.

Facies recorded at this locality include C2.1, C2.2, C2.3, (D2.1), (D2.2), E1.1 and E2.2.

Sandstones are sharp based with graded tops. Reverse grading is noted at the base of some sandstone beds while the tops of others are rippled. Internally, they display complete Bouma (1962) sequences as well as a variety of partial ones, for

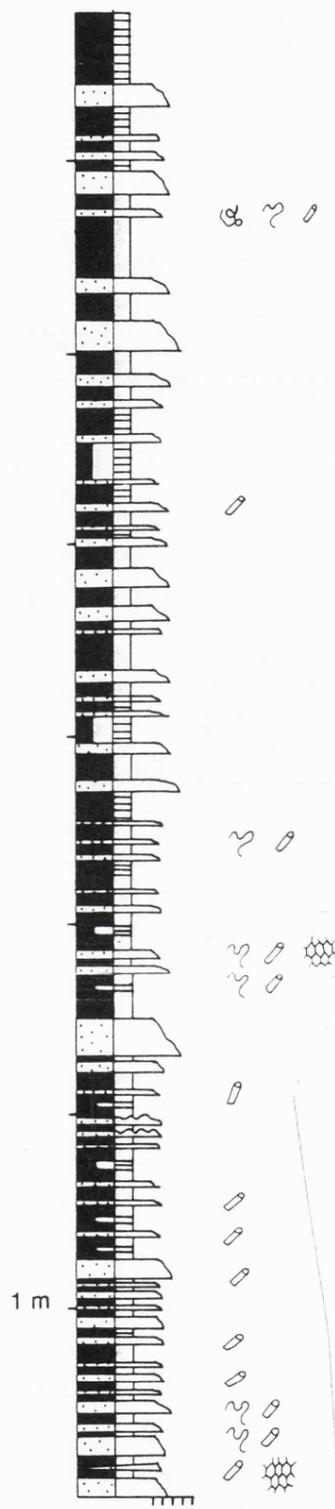


Figure 3.23 Measured section of Aberystwyth Grits Formation, New Quay Harbour (Section 3).

example, Tacde, Tace, Tbcde, Tbce, Tcde and Tce. Tc divisions may be convolute. Some Td divisions show evidence of having been deposited from dilute turbidity currents. A number of the sandstone beds are seen to grade laterally from Tcde to Tce.

Siltstone beds are rare and commonly laterally discontinuous over distances of metres. They are sharp based and internally may be classified as Tde.

Mudstones are both structureless and parallel laminated.

### 3.5 Summary

Each of the formations may be grouped, depending on both the types of sediment facies and their observed arrangement, into one of four depositional environments. These environments, namely distal shelf, sandstone lobes, lobe fringe and slope, will be discussed in more detail in chapter four.

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## CHAPTER 4 - SEDIMENTARY ENVIRONMENTS OF THE LLANDEILO-LLANDOVERY COASTAL SUCCESSION

### Abstract

The sediments of the Llandeilo-Llandovery succession are interpreted as representing distal shelf, sandstone lobe, lobe fringe and slope environments. The turbidite systems developed in the Llandeilo-Llandovery coastal succession within the Welsh Basin as a Type 1 (unchanneled sandstone lobes) system. Lobe development is closely linked to global eustatic sea-level variations although tectonic influences are locally important. The Sea of Okhotsk (Western Pacific Ocean) is a useful modern analogue for the Lower Palaeozoic Welsh Basin.

### 4.1 Tectonic setting and palaeogeography

The following section outlines the tectonic setting of the Welsh Basin and the palaeogeography of the basin from the Cambrian to the late Llandovery.

#### 4.1.1 Tectonic setting of the Welsh Basin

The possible tectonic setting of the Welsh Basin has been the subject of a number of recent reviews (e.g. Holland et al. 1979; Bassett 1984; Woodcock 1984; Siveter et al. 1989). The early Palaeozoic Welsh Basin was a relatively rapidly subsiding area of continental crust separated from the more stable Midland Platform to the south and southeast by an arcuate array of steep (at the surface) faults (Woodcock 1984) (Fig. 4.1). The area is thought to have been the site of a NE-SW elongate, ensialic, back arc, and at times marginal, depositional basin located along a destructive plate margin on the southern side of the Iapetus Ocean from

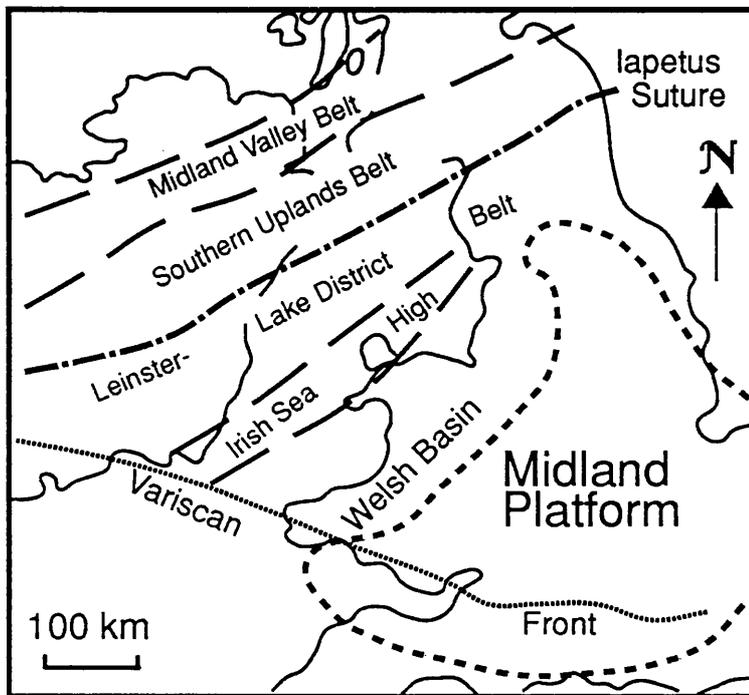


Figure 4.1 Regional and tectonic framework of the Welsh Basin (modified after Ziegler 1984).

the late Precambrian through lower Palaeozoic times.

Southern Britain, including the Welsh Basin, formed part of the southerly continent of Gondwana. During the Ordovician Eastern Avalonia, a microcontinent comprising Britain south of the southern Uplands and much of southeast Ireland, broke away and drifted northwards. This was followed by the closure of Tornquist's Ocean between Eastern Avalonia and Baltica by the late Ordovician (Caradoc-Ashgill) and was coeval with the opening of the Rheic Ocean to the south of Avalonia, a model supported by faunal evidence (Cocks & Forzey 1982; Vannier *et al.* 1989). By the early Silurian the Iapetus Ocean separating Eastern Avalonia and Baltica from Laurentia had effectively closed (i.e. no more oceanic crust) although in the British succession marine seaways persisted until the late Silurian (Pickering *et al.* 1988).

The palaeogeographical position, therefore, of the Welsh Basin together with the varying influences of volcanic, tectonic and eustatic sea-level changes produced a complex facies mosaic throughout the area. A general view, based on facies analysis, is that the basin was deeper in the centre, shallowing to the south (south of the Twyi Lineament) and in the Welsh Borderlands and English Midlands (Woodcock 1984). The Welsh Basin was probably underlain by Precambrian continental crust throughout the early Palaeozoic (Watson & Dunning 1979). It has been suggested that this crust is largely the product of late Precambrian-early Cambrian arc accretion events (Thorpe 1979; Thorpe *et al.* 1984).

Geochemical evidence from the volcanic centres in Wales records a period of late Tremadoc arc volcanism followed by Arenig-Caradoc back-arc extension (Kokelaar *et al.* 1984) with the main arc being situated in the Leinster-Lake District zone to the northwest (Phillips *et al.* 1976). Subduction was towards the south. The recorded changes, both in the geochemistry and the style of volcanism, between the Tremadoc and the mid-Ordovician represent a transition from a volcanic arc to a more marginal basin setting (Bevins *et al.* 1984; Kokelaar 1988).

The Ashgill and Silurian plate-tectonic setting of the basin is difficult to determine due to uncertainties concerning the exact timing of the final closure of Iapetus Ocean and the relative effects (Cocks & Fortey 1982), the extent of strike-slip displacements associated with the Armorica-Laurentia collision (Murphy & Hutton 1986; Soper 1986) and the importance of other convergence zones to the south and northeast (Soper 1986). The eastward migration of depositional centres during the Silurian led Okada & Smith (1980) to suggest a forearc setting for the region involving compressional uplift of the Irish Sea High as a result of southeastward migration of a southeasterly-dipping subduction zone. This subduction pattern could then be related to volcanic rocks of possibly Lower Silurian age at Marloes in south Wales (Ziegler et al. 1969) and at Silurian inliers near Bristol (Hurst et al. 1978; Stillman & Francis 1979). An alternative explanation, however, for the presence of Silurian igneous rocks is suggested by Soper (1986) who relates them to northward subduction to the south of the Welsh Basin, that is the area between the Midland Platform and the Ardennes. A further problem with the forearc model is the lack of evidence for a subduction related accretion zone in the Leinster-Lake District Belt (Woodcock 1984; Kelling et al. 1985). Other models which have been suggested for the Welsh Basin include a passive continental margin setting (Davies & Cave 1976) and a strike-slip continental borderland similar to modern day California (Woodcock 1984a).

The sites and patterns of sedimentary and volcanic activity were affected by the nature and distribution of major structural displacements within the basin (Kokelaar 1988) (Fig. 4.2). Evidence from the Ordovician volcanic centres suggests that they were located above basement faults and involved both subcrustal melting together with admixture of continental crust (Woodcock 1984a). Such a situation is consistent with strong crustal thinning on listric normal faults but is more likely with steep transtensional strike-slip faults which provide a direct passage to the mantle (Woodcock 1984a). The transfer of active extension between graben-related fractures within these deep crustal

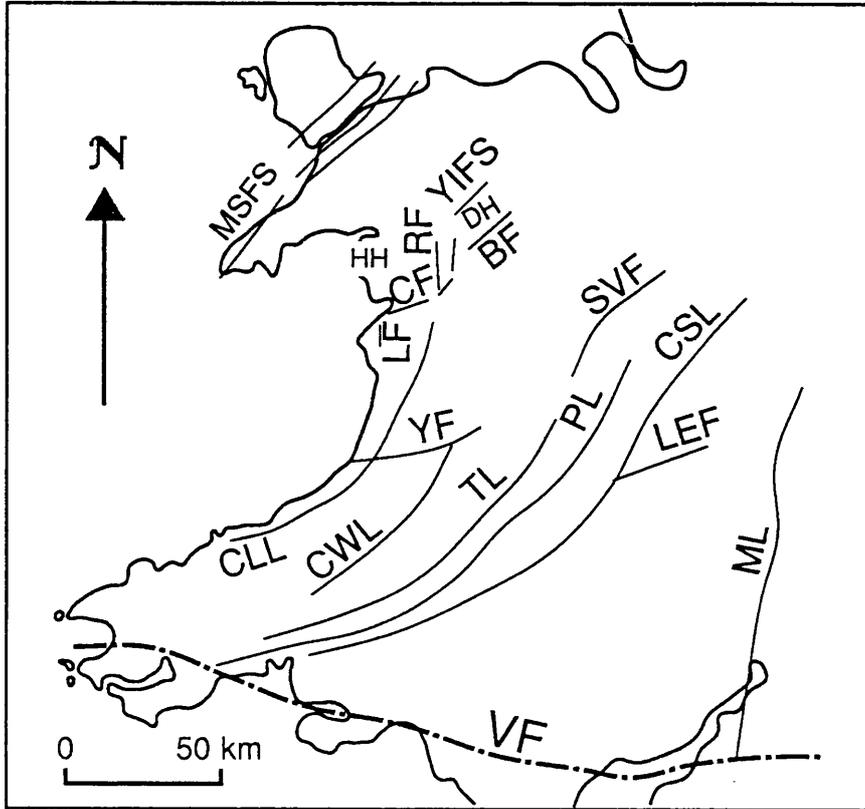


Figure 4.2 The "Variscan Front" and structures inferred to have been active during the early Palaeozoic history of the Welsh Basin: BF, Bala Fault (James & James 1969; Fitches & Campbell 1987); CLL, Corris-Llangranog Lineament (James & James 1969; James 1972; Craig 1985; Craig 1987); CF, Ceunant Fault (Fitches & Campbell 1987); CSL, Church Stretton Lineament (Holland & Lawson 1963); CWL, Central Wales Lineament (Smith 1987a); DH, Derwen Horst (Fitches & Campbell 1987); HH, Harlech Horst (Fitches & Campbell 1987); LEF, Leinthall Earls Fault (Lawson 1973); LF, Llanegryn Fault (Fitches & Campbell 1987); ML, Malvern Line (Hurst *et al.* 1978); MSFS, Menai Straits Fault System (Kokelaar *et al.* 1984); PL, Pontesford Lineament (Woodcock 1984b); RF, Rhobell Fracture (Kokelaar *et al.* 1984; Fitches & Campbell 1987); SVF, Severn Valley Fault (James & James 1969); TL, Twyi Lineament (Smallwood 1986); YIFS, Yspytty Ifan Fault System (Fitches & Campbell 1987); YF, Ystwyth Fault (James & James 1969).

strike-slip duplex systems may also explain the frequent switches of volcanic centres through Ordovician times (Kokelaar 1988).

Deformation, therefore, of the Lower Palaeozoic rocks of Wales involved both strike-slip and dip-slip movements (Woodcock 1984a; Siveter *et al.* 1989). Several of the faults demonstrate basement control and there is evidence of repeated reactivation (James & James 1969; James 1972; Kokelaar *et al.* 1984). The anastomosing pattern of major lineaments may be partly a reflection of strike-slip displacement during oblique accretion episodes (Woodcock 1984a, 1986; Woodcock & Gibbons 1988). Folds are arranged in an arcuate band and tend to be upright with the fold axes verging towards the southeast (Woodcock 1984a).

To the north and east of the Welsh Basin lie the Menai Straits Fault System and the Welsh Borderland Fault System respectively. The former system is a probable terrane boundary which formed as a result of Monian terrane amalgamation. Transcurrent movements associated with this amalgamation may have led to the origin inboard within Eastern Avalonia of the possible terrane boundary of the Welsh Borderland Fault System (Woodcock & Gibbons 1988; Siveter *et al.* 1989). It may also have resulted from later extensional tectonics during the main Cambrian-Ordovician developmental phase of the depositional Welsh Basin. The significant contrasts in pre-Ashgill successions across the northwestern edge of the Welsh Borderland Fault System (Pontesford and Twyi Lineaments rather than Church Stretton Lineament) might imply transcurrent displacement in, or prior to, the Ashgill (Woodcock & Gibbons 1988). The observed facies differences, however, do not demand such movements, and could instead be more easily related to the margin of the basin being fault controlled (Siveter *et al.* 1989).

#### 4.1.2 Palaeogeography of the Welsh Basin

For convenience the basin may be divided into four main

subareas each with its own stratigraphy, and separated from each other by major structural breaks.

(1) Irish Sea Platform - The margins of the platform are unrepresented except for Anglesey and the succession is separated from the rest of the stratigraphic record of the basin by an unconformity below the Llandovery.

(2) Northern Welsh Basin - The Cambrian succession of north Wales comprises alternating sandstones and mudstones. The basal and uppermost sandstone beds show features consistent with a shallow water depositional setting while the intervening ones show some evidence of tractional processes although there is little evidence for a bathyal setting (Crimes 1970; Woodcock 1984a). The Harlech Dome area of northern Wales is the type area for the Cambrian System and there a 3,000 m succession is dominated by comparatively deeper water turbiditic sandstones and mudstones.

The early Ordovician-Caradoc succession records both shallow marine and emergent conditions often with a major volcanic influence (Brenchley 1979; Brenchley & Pickerill 1980). Black shales and phosphorites are widely recognised in the Caradoc and record periods of clastic starved deposition possibly enhanced by high nutrient productivity during eustatic rises in sea level (Leggett 1980). There is a major unconformity between the the Caradoc and Ashgill.

Shallow marine conditions persisted into the late Ashgill as evidenced by channels filled with shallow marine sands which formed during the glacioeustatic regression (Brenchley & Newell 1984). The shallow-water facies pass southwest into sedimentary facies indicative of deeper shelf conditions along a line from the Llyn to Shelve area. Southerly-directed palaeocurrents recorded from Machynlleth (Cave 1979) and Borth (McCann & Pickering 1989) suggest that this slope persisted into Llandovery time.

(3) Midland Platform and Welsh Borderland - Shallow subtidal, intertidal and emergent conditions prevailed here from

Cambrian through Silurian times. The Cambrian succession comprises glauconitic and phosphatic limestones and sandstones which appear above basal Cambrian quartzites in Shropshire. This succession records the early Cambrian radiation of metazoan invertebrates and is coeval with similar successions recorded in the English Midlands (c.f. Siveter et al. 1989). Mud-dominated shallow marine subtidal and intertidal conditions of the middle-upper Cambrian (Crimes 1970) gave way to shallow marine conditions during the Tremadoc (Curtis 1968). More nearshore conditions prevailed during the Caradoc (Hurst 1979).

The early Llandovery transgression left some areas, for example the Longmynd, emergent with the development of shoreline facies and sheltered estuaries (Woodcock 1984a). Continued transgression in the late Llandovery flooded much of the platform. The late mid Llandovery shoreline occupied a position along a north-south Malverns-May Hill line. Lateral supply of clastic sediments through deep channels located at the platform edge into the southern part of the basin was important at this time. Typical channels occur at Rhyader (Kelling & Woollands 1969) and further south at Llandovery (Woodcock & Smallwood 1987).

(4) Southern Welsh Basin - The Cambrian is best developed in the St David's region where the sandstones and muds represent shallow water environments. Indeed, pre-upper Ordovician sedimentary successions along the southern margin of the basin are dominated by shallow water deposits. Bassett (1980) recognises three broad facies belts in the Ordovician of the Welsh Basin which extend along the margins of the Welsh Borderland and suggest that perhaps the southern slope of the Basin was steeper than that of the eastern margin.

Early Ordovician volcanism was particularly important near Llandrindod, Llandeilo, Camarthen and in southwest Dyfed. The Caradoc succession is dominated by black shales although turbidites are recorded from the Cardigan area (James 1975; Allen 1981). Ashgill sequences are shallow marine but northwest of the Twyi Lineament (which represents the basin

edge) deposition is dominated by deeper-water turbiditic clastics derived from the platform to the southeast (James 1972, 1981, 1983).

The late Ordovician regression was associated with the probable emergence of a Precambrian basement area, termed Pretannia (Cope & Bassett 1988), with lagoonal and shallow marine facies being developed along the shoreline. West of the Tywi Lineament the early-mid Llandovery was largely characterised by deposition of graptolitic mudstones and turbiditic sandstones. The sandstones are derived from the eastern edge of the platform (Cave 1979) while the late Llandovery Aberystwyth Grits were derived from the south and southwest and transported axially (Wood & Smith 1959).

#### 4.2 Palaeoflow in the succession

Palaeocurrent indicators are uncommon in the study area due to the combination of concealed bed bases and the original lack of sole structures. Bottom structures, for example flute casts and groove marks, are the dominant indicators of current direction although other features, including current ripples and slump vergence, were also used (Fig. 4.3). Palaeocurrent directions, where present, do show a consistent direction of sediment transport to the north-northeast for much of the period of deposition, except in region to the north of Aberystwyth (see Section 4.2.1).

Of particular interest is the fact that the coarser sediments show consistent palaeocurrent directions while the finer-grained sandstone and siltstone beds show variations from the predominant direction. The coarser-grained sediments (e.g. Newport Sands Formation, Aberystwyth Grits Formation) tend to show only the dominant palaeocurrent directions. Variations from this direction are particularly well exhibited by those palaeocurrents recorded from the Gwbert and Mwnt formations. Craig (1985) interpreted the variations in terms of traditional fan models prograding in the direction of palaeocurrent from a point source to the

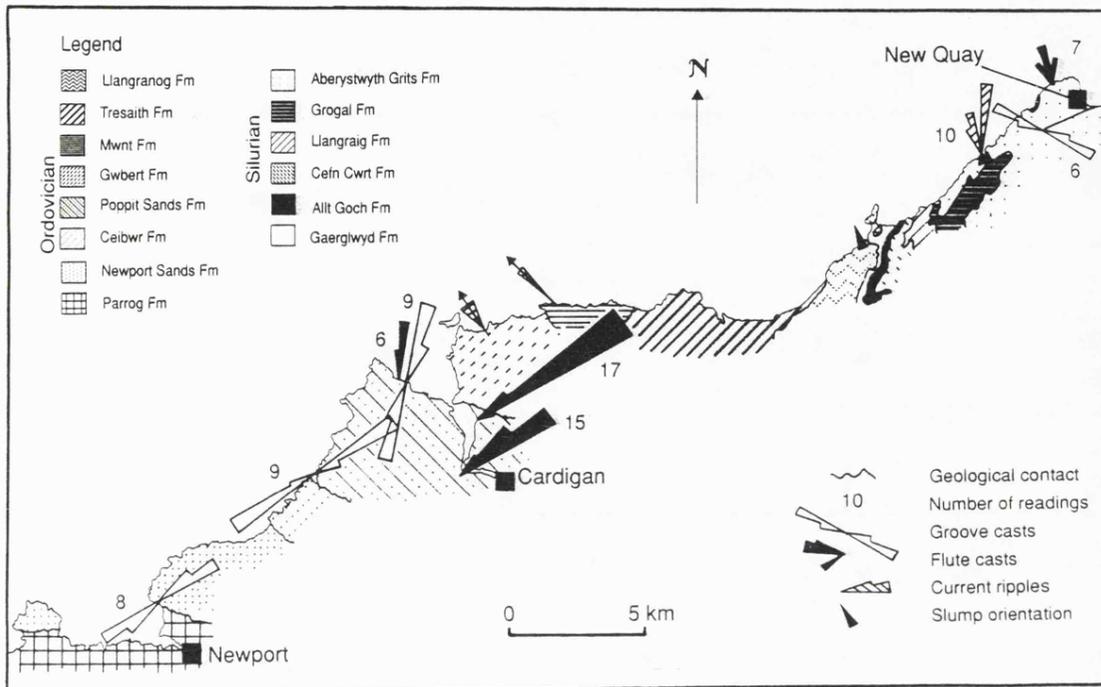


Figure 4.3 Palaeocurrent map of the study area. Palaeocurrent vector means for the Gwbert and Mwnt formations after Craig (1985).

east or northeast of the depositional area. The elongate, fault controlled nature of the Welsh Basin, however, would preclude the development of such fan systems and instead favour the growth of elongate sandstone lobes (see Sections 4.4 & 4.5). The sediments in both the Gwbert and Mwnt formations are typically deposited from dilute turbidity currents and were probably positioned lateral to the main sandstone lobes as lobe-fringe or interlobe deposits. Depending on which way the current was deflected off the sandstone lobe, therefore, the palaeocurrent patterns would be expected to display variations of up to  $90^{\circ}$  in either direction from the main palaeocurrent trend. Other observed variations in palaeoflow are noted in inferred interchannel deposits from the Aberystwyth Grits Formation and will be discussed below (see Section 4.4.3).

#### 4.2.1 Anomalous palaeocurrents to the north of Aberystwyth

Twenty-eight palaeocurrent determinations were made in the Wallog-Borth region, to the north of Aberystwyth, in a section approximately 30m thick which comprises mainly pelagic and hemipelagic mudstones with some intercalated, laterally continuous, fine- to medium-grained sands and silts (Fig. 4.4). The mean palaeocurrent direction is to the south-southeast ( $166^{\circ}$ ) which is at variance with previously reported palaeocurrent directions for the area. The deposits show no evidence of reworking in their upper portions; grading is normal and the cross-lamination is not burrowed. Bioturbation in the form of graphoglyptid burrows, is locally present in other facies and at the base of some of the turbidite deposits. The deposits are, as far as can be determined, primary Tcde and Tce divisions of turbidity currents.

Investigations of the area immediately to the south of Wallog reveals a distinct lack of interdigitation between those deposits showing palaeocurrent directions to the north-northeast. There is no evidence for reflected flows (sensu Hiscott & Pickering 1985). It is not realistic to

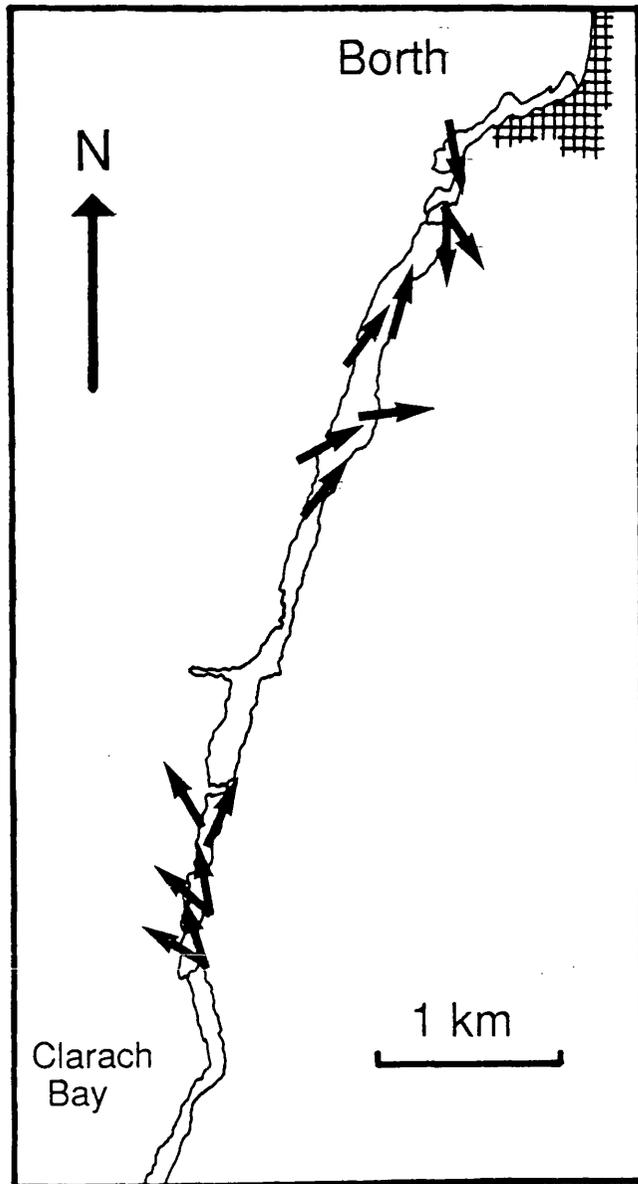


Figure 4.4 Palaeocurrent map of the area between Clarach and Borth.

have totally opposing flow directions ( $140^{\circ}$  apart) of the same age only separated by a lateral distance of 1 km. It is entirely possible that the anomalous palaeocurrent direction may be the result of the presence of a northern structural high and sediment source. The Derwen Horst (Fitches & Campbell 1987) is an ideal candidate which would explain the reported palaeocurrents and also the deflection of the broadly northeasterly palaeocurrents in the Wallog-Borth area (McCann & Pickering in press).

### 4.3 Vertical sequence analysis

The study of ancient turbidite systems may rely, to a great degree, on the interpretation of vertical sequences. The presence or absence of distinctive vertical sequences has been used to recognise and distinguish different depositional environments present on submarine fans. Walker & Mutti (1973), for example, suggested that thickening-upwards sequences represented lobe progradation while upward-thinning sequences formed under conditions of channel infilling. Given that these environments are very different, both in terms of their possible position on a submarine fan and also in terms of associated environments, it follows that more rigorous criteria must be used in their identification rather than the simple presence or absence of vertical sequencing. Such a view was supported by Mutti & Normark (1987) who favoured external geometry for recognition of environments. The reliance on vertical sequences in providing simple answers was also challenged by Hiscott (1981) who noted that asymmetric vertical sequences were uncommon and that aggradation and lateral shifting, rather than simple progradation, dominated submarine fan construction. Such processes would tend to produce random rather than fixed vertical sequences.

Runs tests are a statistical approach to identifying cyclicity in vertical sedimentary sequences. Their method involves the testing of a null hypothesis which states that: the thickness of beds in the stratigraphic section is random.

LOCATION	D	N	U	Uu	Ou	Z	Ho	Max+ve	Max-ve
NS	R	57	35	37.67	3.13	-0.85	Accept	4	3
	S	29	21	19.0	2.2	+0.9	Accept	3	3
PSW	R	92	51	61.0	4.0	-2.5	Reject	5	4
	S	46	33	30.33	2.8	+0.95	Accept	3	2
PSE	R	40	29	26.33	2.6	+1.03	Accept	2	4
	S	20	14	13.0	1.8	+0.56	Accept	2	2
PSR	R	40	28	26.33	2.61	+0.64	Accept	3	3
	S	20	11	13.0	1.8	-1.11	Accept	3	2
ALG	R	18	10	11.67	1.7	-0.98	Accept	3	2
	S	9	5	5.67	1.13	-0.59	Accept	1	4
AG1	R	72	45	47.67	3.53	-0.76	Accept	7	5
	S	36	19	23.67	2.47	-1.89	Accept	6	4
AG2	R	22	15	14.33	1.89	+0.35	Accept	2	3
	S	11	4	7.0	1.28	-2.34	Reject	4	4
AG3a	R	23	13	15.0	1.94	-1.03	Accept	5	3
	S	12	6	7.67	1.35	-1.24	Accept	5	1
AG3b	R	44	26	29.0	2.74	-1.07	Accept	4	6
	S	22	11	14.33	1.89	-1.76	Accept	3	4

KEY: NS - Newport Sands Section 1; PSW - Poppit Sands Section 1a; PSE - Poppit Sands Section 1b; PSR - Poppit Sands Section 2; ALG - Allt Goch Formation Section 1; AG1 - Aberystwyth Grits Formation Section 1; AG2 - Aberystwyth Grits Formation Section 2; AG3a - Aberystwyth Grits Formation Section 3 (bed thickness > 5 cm); AG3b - Aberystwyth Grits Formation Section 3 (bed thickness > 2 cm).

D - Data type (R - Raw bed thickness; S - Bed thickness averaged over two consecutive beds); N - Number of beds in sequence; U - Number of runs; Uu - Expected number of runs; Ou - Standard deviation; Z - Calculated Z score comparable with Z table; Ho - Response to null hypothesis that: Bed thickness in the stratigraphic succession is random; Max +ve & Max -ve - Number of beds in longest upward thickening and upward thinning runs respectively.

Table 4.1 Runs test results for the Welsh coastal succession.

The runs test is a two-sided test where the hypothesis of randomness is rejected if the number of runs is smaller or larger than that predicted. The result of the test is a Z value which is comparable with a table of Z scores.

The test was first used to identify asymmetric cycles in bed thicknesses by Heller & Dickinson (1985). Their method was criticised by Waldron (1986) who questioned some of the assumptions made by the original authors and suggested a number of improvements. These improvements were subsequently incorporated into the method and the revised procedure of Heller & Dickinson (1986) is the one followed herein.

Nine sets of runs tests were attempted and in all but two cases the null hypothesis was accepted, that is, the sequence of bed thicknesses in the measured sections was deemed to be randomly arranged (Table 4.1). For the Poppit Sands section (Section 1a) the hypothesis is rejected for the raw data due to alternating thick- and thin-bedded sandstones giving more runs than predicted. The smoothed data, however, produces a sequence which accepts the hypothesis.

The raw data for a second section (Aberystwyth Grits - Section 2) suggested that the hypothesis should be accepted. When the smoothed data are used, however, there are fewer runs than predicted. Two sequences, each comprising 4 smoothed bed thickness values, one of which is upward thinning while the other is upward thickening are produced by the smoothing process. Visual inspection of the section, however, reveals that no real sequences exist. The vertical sequences suggested by the smoothing process are merely artificial.

#### 4.4 Facies associations and their distributions

Four distinct facies associations are recognised from the study area.

#### 4.4.1 Distal shelf facies (Plate 4.1a)

Description - This facies association is present at the southern end of the study area and is coincident with the Parrog Formation. The dominant facies types present are mudstones, laminated hemipelagites and thin-bedded turbidites. A single ripple showing bidirectional current activity was noted at Parrog.

Interpretation - The muds and turbiditic siltstones and fine-grained sandstones were deposited predominantly below storm-wave base. The presence, however, of the single ripple showing bidirectional current activity does suggest that the area may have been subjected to rare wave-dominated activity.

#### 4.4.2 Slope facies (Plate 4.1b)

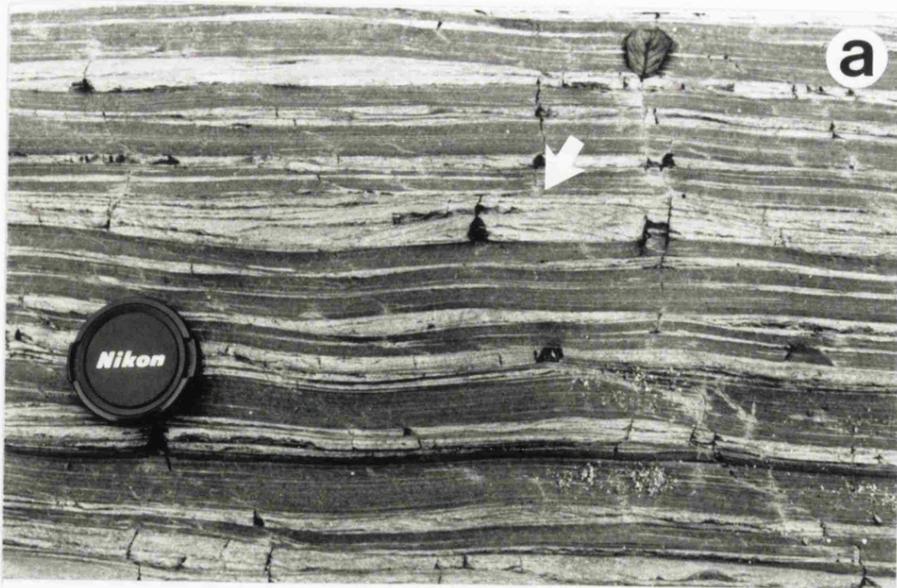
Description - This facies association is present in the middle of the study area and is coincident with parts of the Llangranog Formation. The dominant facies types present are mudstones, siltstone and sandstone turbidites. Coherent folded strata (F2.1) and dislocated, brecciated and balled strata (F2.2) are both present in the section at Llangranog village. Slump sheets up to 6.0 m in thickness are present (Craig 1985). Slump fold orientations indicate a northwesterly dipping slope.

Interpretation - Recognition of the slope facies is predominantly based on the identification of contorted and slumped sediment masses in which the slump fold orientations are consistent. These orientations provide a strike and dip direction for the palaeoslope and the orientations for the slumps in the Llangranog Formation are seen to be perpendicular to the major tectonic lineaments in the area.

Sediment slides are the most widespread type of submarine slope instability. They may be extremely large, for example the Bassein Slide (Indian Ocean) has a total area of 4,000 km<sup>2</sup>, while the Grand Banks Slide (Newfoundland) is estimated

PLATE 4.1

- a. Distal shelf facies, Parrog Formation (Llandeilo), Parrog (SN 04553980). Note the bidirectional current ripple (arrowed) and the predominantly muddy lithology. Lens cap (5.0 cm) for scale.
- b. Slope facies, Llangranog Formation (Ashgill), Llangranog (SN 31095426). Stratigraphic top is to the right. Hammer (30.0 cm long) for scale.
- c. Channel-fill facies association, Aberystwyth Grits Formation (Upper Llandovery), New Quay (SN 38526048). Note the mud-filled scours and lateral discontinuity of the beds. Hammer (30.0 cm long and arrowed) for scale.



as having an area of 27,000 km<sup>2</sup>. Slides are common in areas of tectonic activity and rapid sedimentation. Slopes of only a few degrees ( $< 5^{\circ}$ ) are required and they may even be as low as  $0.5^{\circ}$  (Prior & Coleman 1978; Prior 1984).

Slides may be initiated by either increasing stress or reducing sediment strength. Stress may be increased as a result of rapid sedimentation leading to the input of large amounts of sand to the depositional system. The angle of repose of the sediment mass is thus exceeded and failure may occur, producing a slide. Sediment strength may be reduced by increasing the sedimentation rate such that it exceeds the pore water drainage capacity. The sediment mass would, therefore, become unconsolidated leading to a build up of pore pressures.

The Llangranog Formation was deposited at the end of the Ashgill and thus it appears to be coincident with the late Ashgill regression. Eustatic falls in sea level are a cause of instability in sediment masses. During the Pleistocene, for example, rapid sediment loading on the shelf edges of the Atlantic coast of North America produced slide deposits (Prior & Coleman 1984). It is, therefore, possible that the slumping of the sediments of the Llangranog Formation was related to the drop in sea level. Sediment slides produced by the development of excess pore water pressures can occur at very low angles, and therefore, the slope may have been a relatively gentle one ( $0.5-1^{\circ}$ ).

As noted above the slump direction is perpendicular to the major lineament trend in Wales. Of particular interest in this case is the position of the Corris-Llangranog Lineament immediately adjacent to the slide deposit. Tectonic activity sited along the lineament could have acted as a trigger mechanism for the slide and cannot be excluded as a possible cause.

**4.4.3 Sandstone lobe facies association (Plate 4.1c; Plate 4.2a, b)**

Description - Three subordinate facies are described:

(a) Channel facies (Plate 4.1c) - This facies is well developed at two locations in the study area. At Poppit Sands the facies comprise sandstones and siltstones with minor mudstones. Disorganised gravelly muds are present locally and may be indicative of deposition from either a cohesive debris flow or a slide deposit. Quarry sections to the south of New Quay show very thick coarse sandstones, with frequent amalgamations, which possibly represent depositional channel fills. Overall within the section there is evidence of upward thinning and fining.

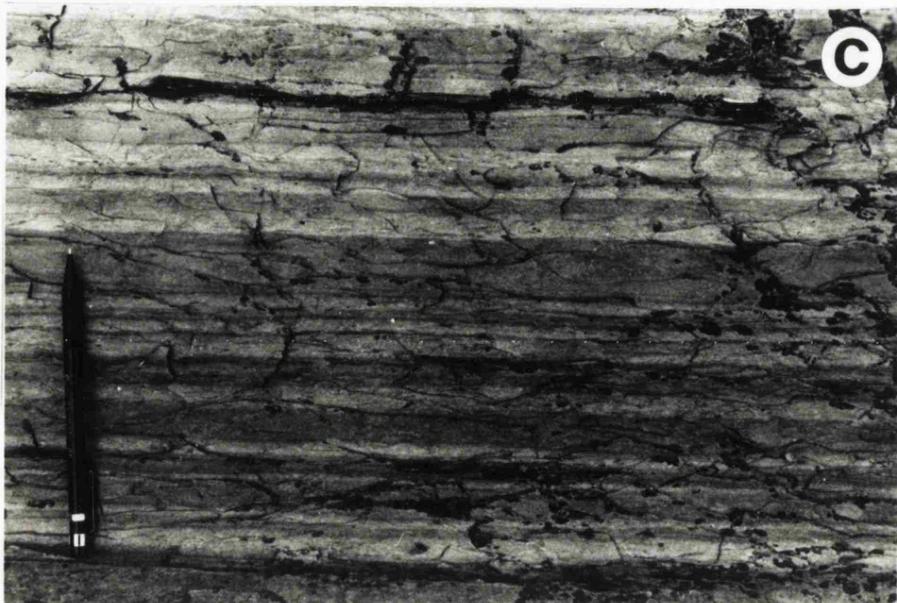
(b) Overbank facies (Plate 4.2a) - The channel margin facies is particularly well developed to the north of New Quay where the exposed sediments are the lateral equivalents of the coarser channel sediments to the south. The sediments are generally fine- to medium-grained and show evidence of fining laterally. Of particular interest is the observation that in a number of the sandstone beds the Bouma (1962) sequences are seen to change laterally from, for example, a T<sub>cd</sub> division to a T<sub>ce</sub> one in a downcurrent direction. This would suggest that these features are the product of waning currents such as would be produced in an overbank wedge. Palaeocurrent directions are also observed to be divergent from the main palaeocurrent trend noted in the adjacent, inferred channel deposits.

(c) Sandstone lobes sensu stricto (Plate 4.2b) - The majority of the sand-sized sediments in the study area are interpreted as having been produced on sandstone lobes sensu stricto (following the definition of Mutti & Normark 1987). The sandstone beds are non-channelised and laterally continuous showing no evidence of lensing or large scale erosional features. The sequences are arranged in apparently random successions with an almost total absence of thinning and fining upwards cycles.

Interpretation - All three facies are intimately related and together may be interpreted as forming sub-environments on

PLATE 4.2

- a. Interchannel facies association, Poppit Sands Formation (Caradoc), Poppit Sands (SN 14604960). Notebook (20.0 cm long) for scale.
- b. Lobe sandstone packet, Newport Sands Formation (Llandeilo), Pwll Côch (SN 06504350). Note the non-channelized, laterally continuous nature of the lobe sandstone packet.
- c. Lobe fringe thin-bedded turbidites, Gwbert Formation (?Caradoc), Gwbert (SN 16185121). Pencil (14.0 cm) for scale.



sandstone lobes. The channels and overbank deposits are subordinate to the more abundant lobe sediments. Shanmugam & Moiola (1988) note the dominant characteristics of submarine fan channels. Based on criteria, for example, the presence of erosive sedimentary features and the abundance of mudstone rip-up clasts, and the inferred mass transport processes operating at the time of deposition, the sediments noted above may be classified as channel deposits.

The sheet-like, laterally continuous, non-channelised nature of the sandstone beds together with the relatively low percentage of amalgamated beds, the presence of only minor scouring and the random arrangement of the bed thicknesses are all features consistent with a lobe interpretation (Pickering 1981; Mutti & Normark 1987).

Lobes are produced either by vertical aggradation (Hiscott 1981) or by basinward progradation (Mutti & Ricci Lucchi 1975) although Shanmugam & Moiola (1988) have suggested that both are responsible. The presence of thickening upwards sequences have been used to distinguish the two forms, with Ricci Lucchi & Valmori (1980) suggesting that the absence of these sequences may indicate that the lobes formed by aggradation. Smaller scale thickening upwards cycles, termed "compensation cycles" (Mutti & Sonnino 1981) have also been reported. The absence of major thinning and thickening upward cycles (as suggested by the vertical sequence analysis) would suggest, therefore, that the lobes in the study area were aggradational lobes. Shanmugam & Moiola (1988), however, note the lack of standard criteria for the recognition of aggradational lobes and it is, perhaps, better not to classify them too rigidly.

#### 4.4.4 Lobe fringe facies association (Plate 4.2c)

Description - The facies comprises parallel-laminated and structureless mudstones intercalated with turbiditic siltstones and rarer sandstones. Laminated hemipelagites are locally abundant in certain areas (Tresaith Formation,

Gaerglwyd Formation). Mudstone beds are laterally continuous while the siltstone and fine-grained sandstones are seen to vary laterally.

Interpretation - These sediments are positioned both at the margins of the lobe sediments and between the major sandstone lobe packets. The hemipelagic sediments represent the background sedimentation of the basin while the silt turbidites show the effect of dilute turbidity currents. The inferred palaeogeographic position of the sediments with respect to the sandstone lobes favours their interpretation as lobe-fringe deposits.

#### 4.5 Palaeogeography of the study area

The turbidite systems developed in the Llandeilo-Llandovery coastal succession within the Welsh Basin may be interpreted as a Type 1 system of Mutti & Normark (1987) and other authors (Nilsen 1980; Mutti 1985). These are also often referred to as high efficiency fans (Mutti 1979) although Shanmugam & Moiola (1985) consider this term to be highly misleading. Idealised Type 1 turbidite systems are dominantly composed of unchanneled sandstone lobes where the lobes may be correlated with erosional channels upstream. They are particularly characteristic of Type C basins where tectonic activity produces and maintains narrow depositional basins. Such narrow basins tend to enhance sediment transport in a similar way to trenches (e.g. in the Alaskan Chugach Terrane trench sediments).

Type 1 systems commonly have only one major site of sediment input, for example a major river or delta complex, and it is entirely probable that such a sediment source was active on the southern margin of the Welsh Basin from Llandeilo-Llandovery times.

Framework grain analysis of the sandstones from the Newport Sands, Poppit Sands, Llangranog and Aberystwyth Grits formations shows evidence of sediment recycling (see Chapter

Five). This would suggest that some of the late Precambrian basement was exposed in the source area. The presence of granophyre and grains indicative of multicyclic origin suggest that this process is a long-lived one. Such a conclusion would be supported by evidence of recycling noted from Arenig-age sediments deposited to the south of the present study area (Traynor 1989).

During periods of high sea level stand, sediment input to a depositional basin is generally reduced with shelf and nearshore areas acting as sites of sediment storage. An exception to this case is when the continental shelf is narrow and the submarine canyon feeding the deep-sea fans extends across the entire shelf almost directly to the river mouth (Heezen et al. 1964; Shanmugam et al. 1985). In these situations sediments are discharged into the canyon, therefore bypassing the shelf, and fan development may proceed even during transgressive phases. The situation in the study area, however, where sediment input is very clearly related to the sea level curve, with pulses of coarser sediment entering the system during periods of low sea-level stand and ceasing as sea levels began to rise would suggest that the shelf area to the south was not as narrow as has been suggested (Bassett 1980). Certainly the shelf area would have needed to be large enough to have acted as a sediment trap and store during periods of higher sea level.

Consistent palaeocurrent directions and elongate sediment bodies within a basin can be produced in one of two ways. The first would involve a relatively steep slope which would therefore provide increased momentum to the mass flow deposits. There is, however, no evidence for a steep slope to the southern end of the Welsh Basin during the period of deposition within the study area. The second method would involve structural control of the basin to produce elongate basins with fault bounded margins. Sediment would, therefore, be channeled as if in a trough. This latter explanation is the one favoured in the present situation.

Palaeocurrents reflect confinement of flows showing a very

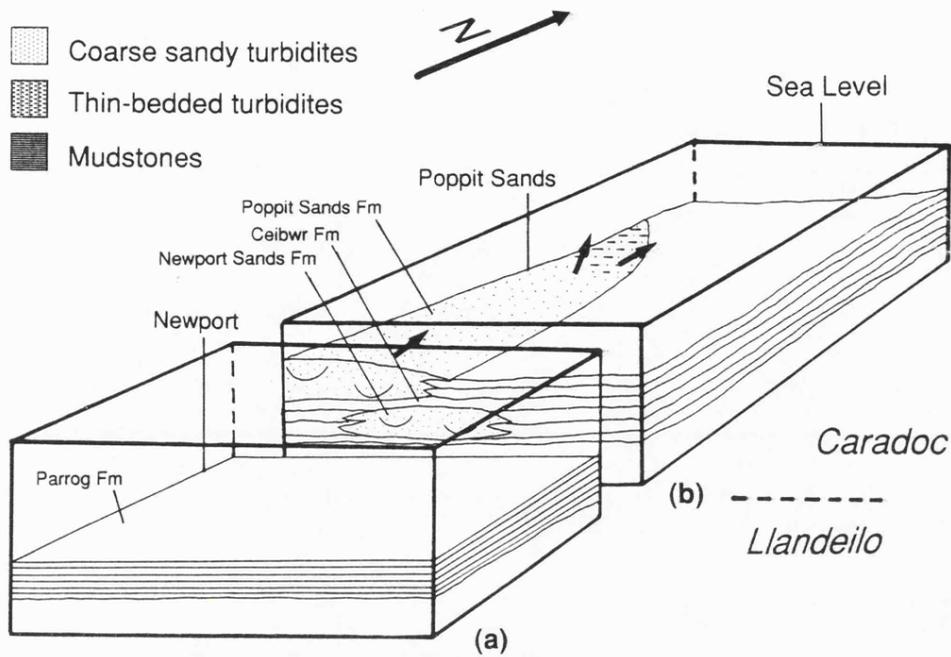


Figure 4.5 Block diagram showing a facies reconstruction for the study area during the Llandeilo and Caradoc (see text for details).

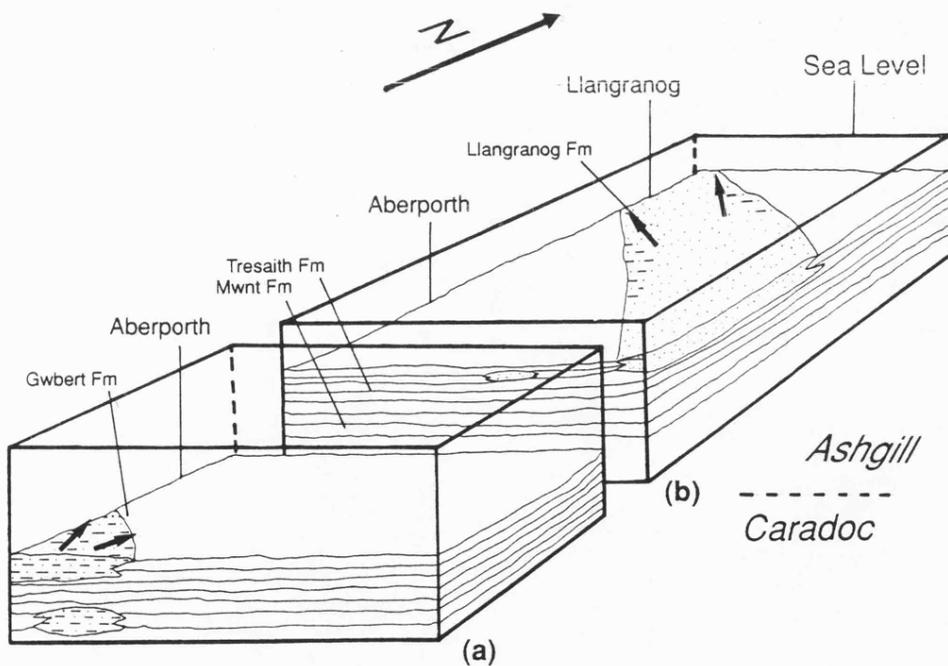


Figure 4.6 Block diagram showing a facies reconstruction for the study area during the Caradoc and Ashgill (see text for details).

consistent flow to the north-northeast for much of the period of deposition in the study area. Palaeocurrent directions are very consistent through the depositional area both areally and stratigraphically. The main trend of the palaeocurrent direction parallels the major lineament trends within the southern end of the Welsh Basin, particularly the Central Wales Lineament (formerly the Central Wales Syncline), the Twyi Lineament and the Corris-Llangranog Lineament. Previous work has shown that related lineaments, for example the Pontesford and Church Stretton lineaments, have greatly influenced sediment distribution patterns within the Welsh Basin (Dimberline & Woodcock 1987; Smith 1987b) and therefore it is probable that the lineaments adjacent to the study area exercised similar controls.

Thin-bedded turbidites in the Gwbert and Mwnt formations show palaeocurrent directions which are at variance to the general trend of those shown by the coarse sandstone lobes. These divergence patterns may be related to deflection of dilute turbidity currents off minor topographic highs, for example, sandstone lobes. An alternative interpretation, however, may be that the thin-bedded mudstone turbidites of these (and possibly other, stratigraphically younger, formations) may have been derived from slumping of minor sediment masses at the slopes created by the lineaments and the redeposition of this sediment further into the basin.

#### 4.5.1 Llandeilo-Caradoc palaeogeography

The lack of a firm biostratigraphy in the southern part of the study area makes it difficult to determine the exact sequence of events within the region. The oldest formation (Parrog Formation) is interpreted as having been deposited in a distal shelf environment (Fig. 4.5a). The abundance of laminated hemipelagites and their geochemistry (see Chapter Six) suggests that the environment was anoxic, or at least dysaerobic. The sea-level curve records a rise during the Llandeilo and into the Caradoc (McKerrow 1979; Leggett et al. 1981). Two major sandstone lobe turbidite systems

(Newport Sands Formation, Poppit Sands Formation), sourced to the southwest of the area, were deposited during the Nemagraptus gracilis Zone (Llandeilo) and during the Dicranograptus clingani Zone (Caradoc) (Fig. 4.5b). This would, therefore, suggest that:

(1). Deposition was related, not to the low sea-level stand, but to the onset of the transgressive phase.

(2). The controlling mechanism was not the eustatic rise in sea-level, but perhaps some local tectonic influence (possibly related to volcanic activity in north Wales).

(3). Deposition of the sandstone lobes was related to a low sea-level stand which continued into the Caradoc; the rise in sea level did not occur in the study area until after deposition of the Poppit Sands Formation lobe system.

The intervening Ceibwr Formation may represent a change in depocentre or else a change in the type of sediment being carried into the depositional basin.

#### 4.5.2 Caradoc-Ashgill palaeogeography

The Poppit Sands Formation contains a graptolite fauna indicating a Caradoc age for the area. The date, however, is not any more exact and there is, therefore, a problem dating the succession immediately north of the the Cardigan area.

Much of the sediment in the region belongs to the lobe-fringe facies and, therefore, may have been deposited at the margins of coeval sandstone lobes (Fig 4.6a). Much of the sediment, however, is likely to have been deposited during periods of higher sea-level stands when input to the depositional basin would have been reduced. During much of this period the depositional environment was a relatively quiet one with only dilute turbidity currents, possibly related to storms on the shelf, bringing sediment into the area.

Towards the end of the Ashgill there was a major fall in sea level. As noted elsewhere, this regression was linked to glaciation in Gondwana (Crowell 1978; Brenchley & Newell 1984; Caputo & Crowell 1985). The fall in sea level is recorded by the development of sandstone lobes within the Llangranog Formation (Fig. 4.6b), particularly the Penbryn Sands Member (Tata 1987). The second sharp fall in sea level, at the end of the Ashgill, may have been responsible for the sediment slide observed in the Traeth Bach Member of the Llangranog Formation (see Section 4.2.2).

#### 4.5.3 Early-Middle Llandovery palaeogeography

Following deglaciation in Gondwana there was a major eustatic transgressive event. Thus the supply of coarse sediment ceased in the study area, being replaced by the background sedimentation of the Gaerglwyd Formation (Fig. 4.7). Rare, laterally continuous coarse sandstone beds within this formation may be evidence of reduced lobe activity within the area.

The growth of a small sandstone lobe turbidite system (Allt Goch Formation) may have been as a result of local tectonic activity (associated with the Skomer Volcanic Series) rather than sea level changes. Much of the subsequent sedimentary history of the period is dominated by mudstones (Cefn Cwrt Formation, Llangraig Formation). The deposition of the Grogal Sandstone Formation, a small turbiditic flow, immediately prior to that of the Aberystwyth Grits Formation, a distinctly larger turbiditic system, suggests that in this particular instance eustatic sea-level changes within the depositional area were the dominant control on sediment input (as outlined by Mutti & Normark (1987)).

The Grogal Sandstone Formation shows evidence of reworking within some of the sandstone beds. According to Anketell & Lovell (1976) the formation may be interpreted as a contourite deposit. They use the criteria of Bouma (1973) to differentiate turbidites from contourites. These

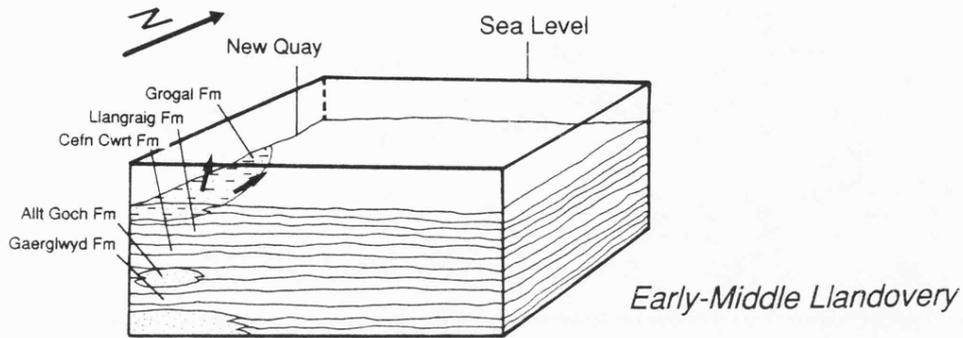


Figure 4.7 Block diagram showing a facies reconstruction for the study area during the Early- Middle Llandovery (see text for details).

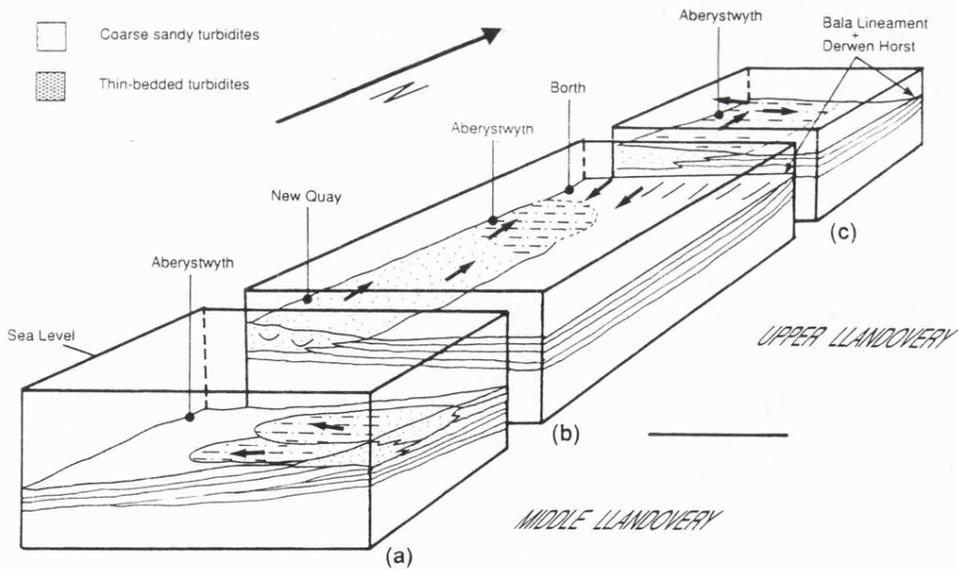


Figure 4.8 Block diagram showing a facies reconstruction for the study area during the Middle-Late Llandovery (see text for details).

characteristics, however, are not wholly diagnostic, and many other factors can be invoked to explain reworking of bed tops rather than geostrophic currents. Another of their arguments in support of contourites was their observation that palaeocurrents between the bed bases and tops varied by  $90^{\circ}$ . Palaeocurrents recorded as part of the present study, however, tend to be much more variable, again supporting the idea of reworking of the bed tops, but perhaps not by geostrophic contour currents.

#### 4.5.4 Late Llandovery palaeogeography

The Late Llandovery is dominated by the deposition of the Aberystwyth Grits Formation, a sandy turbidite system sourced from the south, which accumulated in the axis of the basin and passed laterally and gradationally northwards into the Borth Mudstones Formation (see Cave & Hains 1986 for detailed description of this formation). Deposition of the Derwenlas Formation (see Cave & Hains 1986 for detailed description) from point sources to the east during the middle Llandovery (Fig. 4.8a) was followed by a change in provenance from east to south for the deposition of the Aberystwyth Grits Formation (Fig. 4.8b). Coeval with the initial progradation of the turbidite system of the Aberystwyth Grits Formation there was deposition of the muddier Borth Mudstones Formation by sediments probably derived from a structural high situated to the north and controlled by the Bala Lineament. Parts of the Borth Mudstones Formation farther east and northeast may have been derived from dilute turbidity currents sourced from the Aberystwyth Grits Formation fan system. The final stage involved progradation of the Aberystwyth Grits Formation turbidite system over the muddier Borth Mudstones Formation, the slewing round of turbidity currents towards the northwest and northeast as a result of deflection against the submarine slope parallel to the Bala Lineament (Fig. 4.8c).

#### 4.6 Factors controlling turbidite system development

The development of turbidite systems is controlled by a variety of factors which are interrelated and can be broadly grouped into sea-level changes, tectonic setting and sediment source (Stow et al. 1985). The complexity of these inter-related variables will determine the style of fan growth and the distribution, both vertically and laterally, of sedimentary facies within a deep-water turbidite system.

#### 4.6.2 Eustatic changes in sea level

Global sea-level variations have been interpreted as the primary control on turbidite system development (Shanmugam & Moiola 1982, 1984, 1988; Shanmugam et al. 1985). Short-term, rapid sea-level variations are related to glacial periods in higher latitudes, while longer-term, gradual variations are controlled by factors such as changes in mid-oceanic ridge spreading rates and continental margin subsidence (Shanmugam & Moiola 1988). Shoreline sediment sources, for example rivers and deltas, may have direct access to deep basins during periods of low sea-level stand, given that large areas of the continental shelf will become emergent. Relatively large volumes of terrigenous sediment, therefore, can be transported to the deep sea via turbidity currents and related mass movement flows. During high sea-level stands sediment input is greatly reduced since access to the outer shelf and continental slope environments is restricted. The locus of deposition, therefore, switches to more nearshore environments, for example, barrier island and deltaic complexes with the turbidite systems becoming dormant. Deposition of hemipelagic sediments, however, continues.

The Llandeilo-Llandovery age sediments were deposited during a period of major eustatic changes in sea level (Fig. 4.9). These changes were related to periods of glacial activity to the south in Gondwana. Evidence of glacial activity is discussed by Brenchley & Newell (1984). Using the sea-level curve for the area (Leggett et al. 1981), major inputs of coarse sediment may be related to periods of low sea level

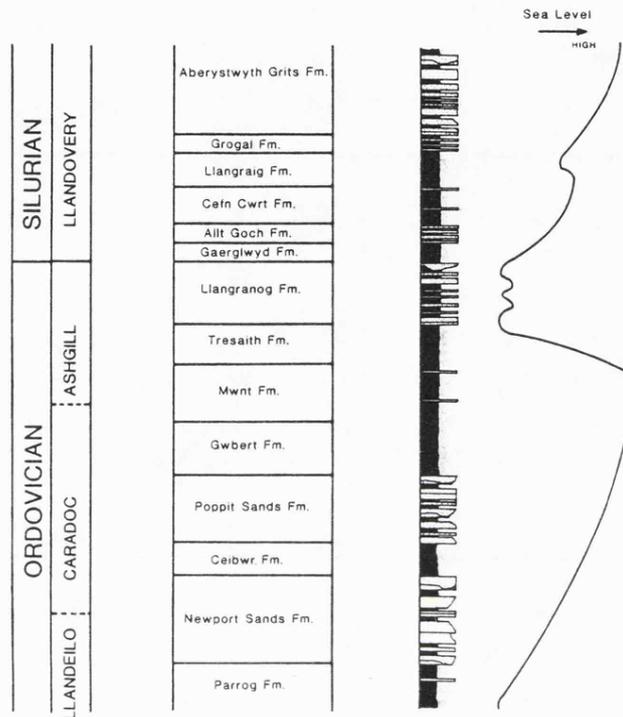


Figure 4.9 Schematic stratigraphic column of the study area relating the lithological divisions to eustatic sea-level changes. Sea-level curve after McKerrow (1979) and Leggett et al. (1981).

stand or periods when sea levels were rising or falling (as discussed earlier). Rising of sea level would have led to abandonment of the turbidite systems and their resultant draping in mudstones and dilute turbidites, some of which may have been related to storm surges on the shelf.

#### 4.6.2 Tectonic setting

The tectonic setting of a basin is an extremely important factor in determining both the type and also the distribution of the turbidite system formed. The tectonic setting will determine the basin shape and size, sea floor and continental shelf gradients, the rates of uplift of the basin margins and the style and frequency of seismic activity and faulting. Mutti & Normark (1987) suggested that submarine fan morphology, internal structure and facies associations present were primarily determined by the long term stability of the basin and the volumes of sediment supplied (see below). They subdivided turbiditic basins into four types each characterised by the type of crust underlying the basin together with the intensity of tectonic activity. The Welsh Basin, during the Llandeilo-Llandovery period, was floored by continental crust and was subjected to some fault control along the basin margins. It may, therefore, be classified, following the nomenclature of Mutti & Normark (1987), as a Type C basin.

The influence of basin shape is evident in the elongate morphology of some of the sandstone bodies present in the basin, for example the Aberystwyth Grits Formation (McCann & Pickering 1989) and the Penbryn Sands Member of the Llangranog Formation (Tata 1987). Similar elongate bodies of sediment have been recorded in the Welsh Basin by Smith (1987) for Monograptus griestoniensis Zone turbidites and by Dimberline (1987) for Wenlock turbidites. The elongate basin shape is further documented by palaeocurrent dispersal patterns and also in the case of Smith (1987) by inferred onlap relations.

Fortey & Cocks (1986) note that the lithological and faunal changes in South Wales are in harmony with the global eustatic curve and suggest that it is not necessary to invoke a tectonic influence. As noted above, however, some of the sandstone lobes were formed during periods of transgression rather than at low sea-level stands. Thus tectonic activity may also have been, at least partially, responsible for the deposition of the Llandeilo and Caradoc turbidite systems (Newport Sands Formation, Poppit Sands Formation). Volcanic activity was pronounced in North Wales during the Llandeilo and Caradoc and may have produced earthquakes which would have allowed formation of turbidite systems to the south. The deposition of the Allt Goch Formation may also have had a tectonic influence (see Section 4.5.3). The only other explanation is that sediment input was at a maximum during the changes in sea-level rather than merely at the point of lowest sea level stand. This would accord with the view of Weaver & Kuijpers (1983)

#### 4.6.3 Sediment source

Sediment composition and grain size, together with the volume and rate at which sediments are supplied to an area, are of major importance in determining the morphology of a turbidite system. Medium to high volumes of mud-rich sediment will produce elongate fans with distal sand depocentres whereas lower volume, mud-poor sediment sources will result in small radial systems or, if there are multiple sources, slope aprons (Stow et al. 1985).

As noted earlier the Welsh Basin was a Type C basin during the period of deposition. These basins are formed on continental crust and are characterised by their relatively large and long-lived sediment supply (Mutti & Normark 1987). Sediment supply, therefore, was an extremely important control on the morphology of the turbidite system.

The extensive sand lobes produced in the study area, particularly the Aberystwyth Grits Formation lobe, are

comparable with modern medium-sized turbidite systems fed by large rivers or delta systems (Stow et al. 1985). The sediment source is dominantly mud rich with both the volume and the composition of material supplied to the basin varying with relative sea level. The coarsest material is supplied during periods of low sea level stands.

#### 4.7 The Sea of Okhotsk as a modern analogue for the Welsh Basin

The geological and geomorphological setting of the Sea of Okhotsk allows parallels to be drawn with the Lower Palaeozoic Welsh Basin.

##### 4.7.1 The Sea of Okhotsk

The Sea of Okhotsk is 1500 km long by 850 km wide at its broadest point and is situated on the eastern coast of the USSR between latitudes 44–60<sup>0</sup>N (Fig. 4.10a). It is bounded on the northern side by the landmass of Russia while the eastern and western boundaries are occupied by the Kamchatka Peninsula and Sakhalin Island respectively. The area may be subdivided into three main geomorphic elements based on their relative locations. The northern and central parts of the Sea of Okhotsk form part of an epi-Mesozoic platform built up of deformed geosynclinal rocks ranging in age from Precambrian to Cretaceous and covered by relatively undeformed Upper Paleogene and Neogene rocks (Gnibidenko 1985). These subdivisions correspond to the shelf and continental slope facies respectively. The southern part of the Sea of Okhotsk, the Yuzhno (South) Okhotsk basin is a fault-bounded back-arc basin positioned to the north of the Kuril Islands Ridge, which stretches for over 1,200 km and rises 3.5–4.0 km above the floor of the Basin, and the Kuril-Kamchatka Trench.

The simple geomorphological division, into shelf, slope and deep-marine basin areas, is complicated by the presence of

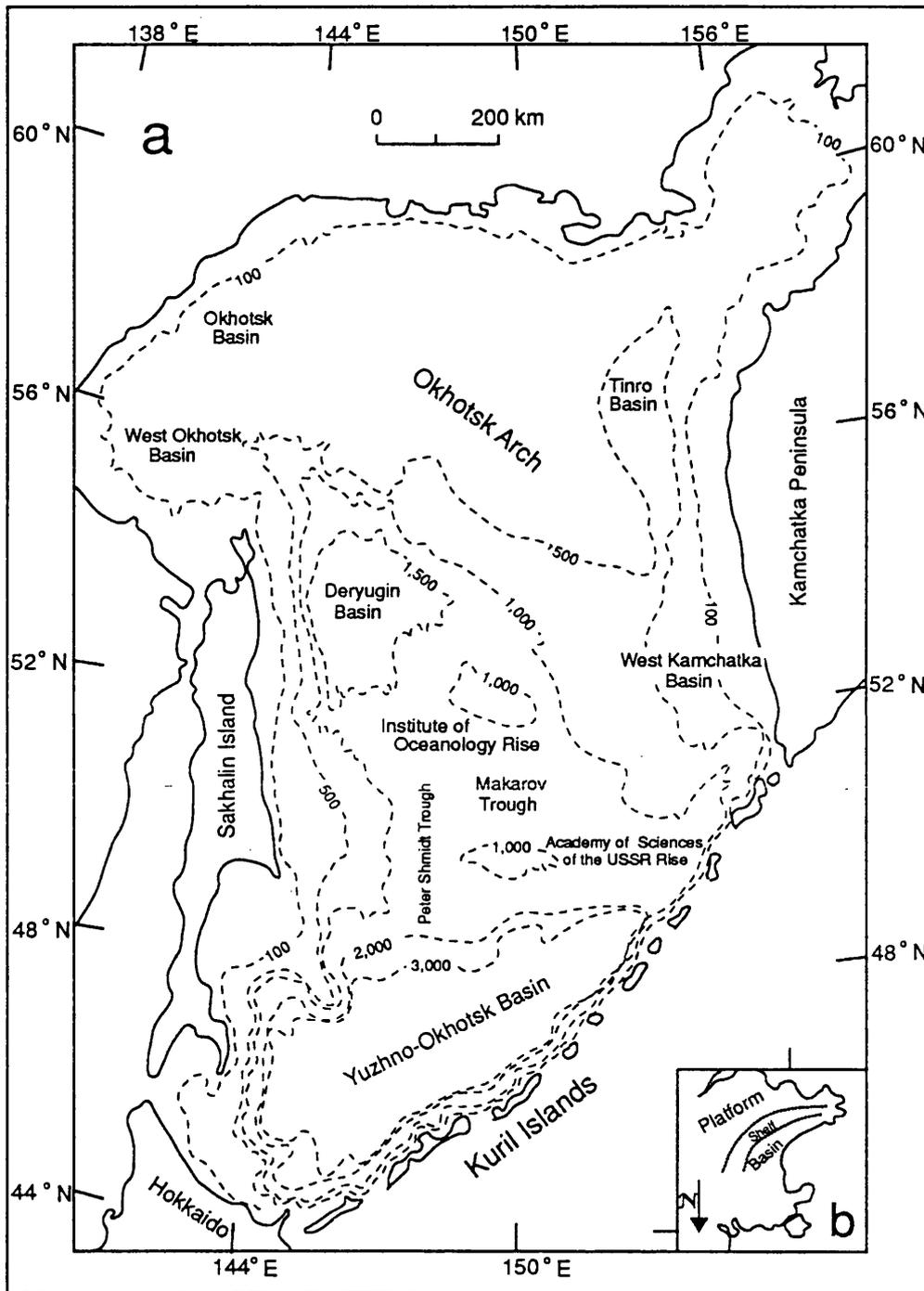


Figure 4.10 Topographic map of the Sea of Okhotsk, eastern USSR and the Lower Palaeozoic Welsh Basin: (a). Topography of the Sea of Okhotsk showing the relative positions of the shelf, slope and trench. All major basins are marked (after Gribedenko 1985). Measurements are in metres. (b). The Lower Palaeozoic Welsh Basin drawn to the same scale as the Sea of Okhotsk and oriented south-north.

smaller basins, for example the Deryugin Basin, and a number of submarine highs such as the Academy of Science of the USSR Rise and the Institute of Oceanology Rise (Gbibidenko 1985). Many of the submarine highs were subaerial during periods of glacio-eustatic low sea-level although it is probable that there was an additional tectonic influence. The glacio-eustatic sea level falls also exposed large areas of the North Okhotsk Shelf (Kulakov 1973).

There are also a number of deep-marine troughs which occur in the northern and central areas of the Sea of Okhotsk. Many of these, for example the Tinro and West Kamchatka Basins, are narrow and elongate parallel to the basin margins. Others, for example the Makarov Trough and Peter Schmidt Trough located in the central area, trend in an east-west direction.

The sedimentary cover rests on irregular basement over much of the Sea of Okhotsk and in some areas it may be traced onshore and thus correlated with Upper Paleogene and Neogene-Quaternary successions (Gnibidenko 1985). In the northern and western parts of the Yuzhno Okhotsk Basin the sedimentary cover on the continental slope has been subjected to faulting, folding, large scale erosion, with associated landslides, and dissection by submarine canyons.

#### 4.7.2 The Welsh Basin and the Sea of Okhotsk

The Welsh Basin is certainly smaller than the Sea of Okhotsk (Fig 4.10b). A broader tectonic setting for the Welsh Basin, however, would include the Leinster-Lake District volcanic ridge to the north. When this is incorporated the comparison is seen to be more pronounced. The Welsh Basin most closely resembles parts of the northern Sea of Okhotsk. The presence of elongate basins, which parallel the coastal configuration (and major lineament trends), acting as sites of active sedimentation with longitudinal and possible lateral input are analogous to similar deep basins, for example the Atlasov Trough or the Deryugin Basin, or some of

the smaller basins to the west of the Okhotsk Arch, in the northern Sea of Okhotsk.

The elongate troughs would act as sites of active deposition during times of maximum sediment input. Such a suggestion is supported by the shape of the sandstone lobes in the study area particularly the Aberystwyth Grits Formation and the Penbryn Sands Member of the Llangranog Formation. Both of these deposits are elongate and show evidence of decreasing current activity towards the north.

Between the troughs the ridges and swells would be areas of background mudstone deposition, some of which may have been distal equivalents of the sandstone lobes when terrigenous input was at its greatest. Redeposition of sediment from some of these low relief structural highs may have given rise to the variable palaeocurrents observed in some of the finer-grained sediments of the study area.

Lack of thorough circulation, and resultant low oxygenation, in some elongate basins (or parts thereof if they are separated up by small highs as in the Sea of Okhotsk) may be a possible interpretation for the presence of hemipelagic sediments in the otherwise aerated Tresaith Formation. It should, however, be noted that this may also be interpreted in terms of the late Ashgill fall in sea level.

The Welsh Basin of the study area was fed by major river systems located to the south, and east of the region during the period of deposition. The Sea of Okhotsk, however, is not fed by any major rivers and this is a major difference between the two areas. Despite this the Sea of Okhotsk does provide, at least geomorphologically, a reasonable analogy for the Lower Palaeozoic Welsh Basin.

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## CHAPTER FIVE - PETROLOGICAL AND GEOCHEMICAL DETERMINATION OF PROVENANCE IN THE SOUTHERN WELSH BASIN

### Abstract

The provenance of Llandeilo to upper Llandovery sediments from the Welsh Basin has been investigated by petrographic and geochemical methods. Sandstone-dominated formations were deposited from high-concentration turbidity currents. Petrographic data suggest derivation from a non-collisional recycled or cratonic setting. This is broadly confirmed by analysis of microconglomerate lithic fragments and the major element geochemistry. Mudstone-dominated formations were deposited from dilute turbidity currents and as hemipelagic sediments. The mudstone geochemistry indicates a tectonic setting transitional between an active continental margin and a passive margin. This study demonstrates that neither set of analytical methods are individually adequate for provenance reconstruction and it is advisable to use a variety of techniques for greater confidence in interpretation.

### 5.1 Introduction

In recent years a number of detrital modal discriminants aimed at the determination of tectonic setting of ancient basins have been developed (e.g. Crook 1974; Dickinson & Suczek 1979). These have been complemented by similar work on modern sediments of known plate tectonic setting (e.g. Valloni & Maynard 1981; Potter 1986). Both are normally restricted to sandstones, although, mudstone geochemistry is increasingly seen as an important area especially by the petroleum industry. The Welsh Basin provides an opportunity to test the use of petrographical and geochemical discriminators against presently accepted tectonic models for the area, which are derived from geochemical studies of volcanic rocks (e.g. Bevins et al. 1984, Kokelaar et al.

1984a).

## 5.2 Tectonic history of the Welsh Basin

The Lower Palaeozoic sediments of the Welsh Basin form part of the Eastern Avalonia terrane and overlie a late Precambrian age lower continental crust comprising calc-alkaline rocks (Watson & Dunning 1979; Thorpe 1979). The early Palaeozoic Welsh Basin was a relatively rapidly subsiding area of continental crust separated from the more stable Midland Platform by an arcuate array of steep (at the surface) faults (Woodcock 1984). The area was the site of a NE-SW oriented, elongate, fault bounded, backarc or marginal basin located along a destructive plate margin on the southern side of the Iapetus Ocean.

Geochemical evidence from the volcanic centres in Wales record a period of late Tremadoc arc volcanism followed by Arenig-Caradoc backarc extension (Kokelaar *et al.* 1984a) with the main arc being situated to the northwest in the Leinster-Lake District Zone. The recorded changes, both in the geochemistry and the style of volcanism, between the Tremadoc and mid-Ordovician represent a transition from a volcanic arc to a more marginal basin setting (Bevins *et al.* 1984; Kokelaar 1988). The detailed Ashgill and Silurian setting of the basin is uncertain, although a number of models have been proposed, including a passive continental margin (Davies & Cave 1976), a forearc (Okada & Smith 1980) and a strike-slip continental borderland, similar to present day California (Woodcock 1984).

## 5.3 Age, Previous Work and Depositional Setting

The study area is located on the west coast of Wales (Fig. 5.1) and comprises a succession of 14 formations (Fig. 5.2) which range in age from Llandeilo (Nemagraptus gracilis Zone) to upper Llandovery (Monograptus turriculatus Zone) spanning a cumulative period of 30-35 Ma (Harland *et al.* 1982). The

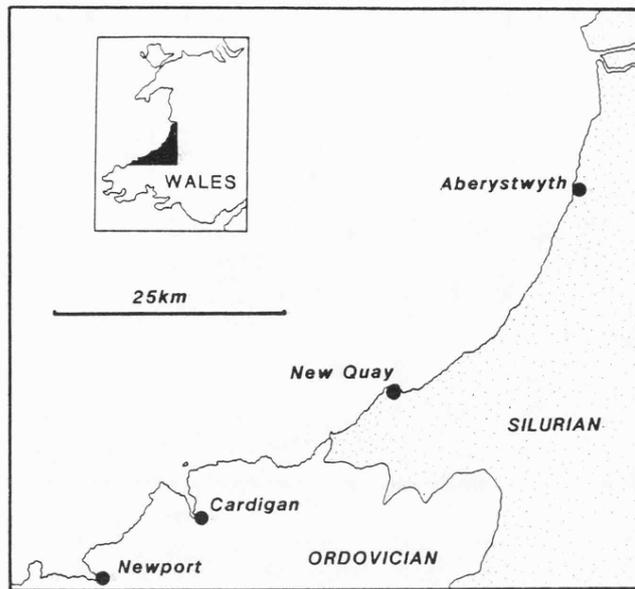


Figure 5.1 Location and simplified geological map of the study area.

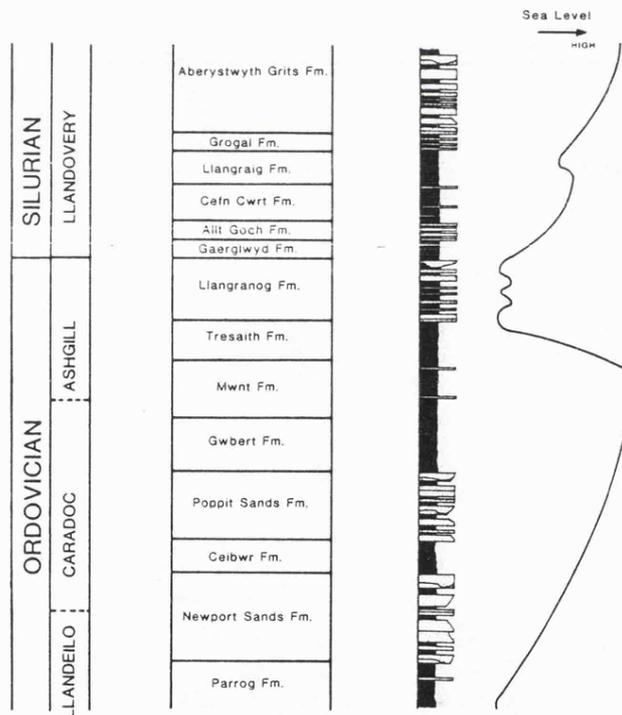


Figure 5.2 Generalised stratigraphic column of the study area relating the lithological divisions to eustatic sea-level changes. Sea level curve after Leggett *et al.* (1981).

formations are identified on the basis of the dominant lithology. There are few studies of the petrographical and geochemical composition of the sediments from this part of the Welsh Basin, apart from work by Bjørlykke (1971), James (1971, 1981), Keeping (1881), Smith (1956) and Wood & Smith (1959).

Sandstone-dominated formations (e.g. Newport Sands Formation, Aberystwyth Grits Formation) were deposited mainly from high-density turbidity currents and, locally, slide deposits whereas mudstone-dominated formations (e.g. Tresaith Formation, Gwbert Formation) were deposited from low-density turbidity currents. Hemipelagites are locally abundant in some of the mudstone-dominated formations (e.g. Tresaith Formation) and predominate in others (e.g. Gaerglwyd Formation, Parrog Formation). The controls on sedimentation appear to have been dominantly eustatic although tectonic influence was also important.

Palaeocurrent evidence reveals that the source area was to the S/SW and sediments were transported parallel to the NW-SE-oriented basin axis. Much of this southerly area was subjected to volcanic activity, some of it quite extensive, in the period immediately preceding the onset of deposition in the study area (Allen 1982; Bevins et al. 1984, 1989; Kokelaar et al. 1984a, b, 1985; Thorpe et al. 1989).

The sediments of the area represent a series of "Type 1" turbidite systems (sensu Mutti & Normark 1987) which are dominantly composed of unchannelled sandstone lobes. They are particularly characteristic of basins where tectonic activity produces and maintains narrow depositional basins (Mutti & Normark 1987). Type 1 systems commonly have only one major site of sediment input, for example a major river or delta complex, and it is probable that such a sediment source was active on the southern margin of the Welsh Basin from Llandeilo-Llandovery times. The large volume of Ashgill and Llandovery deep-marine sediment in the Welsh Basin (and associated areas) suggests that sediments were being eroded from a large landmass to the east (Baltica) (Pickering 1989).

#### 5.4 Sandstone Petrography and Framework Modes

The petrography of the Newport Sands, Poppit Sands, Llangranog and Aberystwyth Grits formations was examined in detail using the Gazzi-Dickinson point-counting technique (see Ingersoll et al. 1984). One hundred and forty six samples were analysed, counting 200-500 points per thin section. All thin sections are stored in the National Museum of Wales (NMW89.13G.T1-161).

Excluding the matrix, which varies between the formations from Newport Sands (15.83%), Poppit Sands (19.33%), Llangranog (16.89%) and Aberystwyth Grits (19.71%), and the patchy calcite cements (1-4%) found in the Llangranog and Aberystwyth Grits formations, the rocks are made up of three main constituents (Table 5.1):

Quartz (Q) - Monocrystalline quartz grains (Qm) are commonly clear although they may be embayed or contain vacuole rims indicative of a volcanic origin (e.g. Newport Sands Formation, Poppit Sands Formation, Aberystwyth Grits Formation). Strained quartz crystals are most common; the lack of any common orientation to the strain shadows suggests that they were strained in the source area. Inclusions within the monocrystalline quartz grains are common particularly muscovite (sericite), zircon and tourmaline. Polycrystalline quartz (Qp) is either chert or strained quartz with sub-grains bounded by crystal faces. Contacts between the sub-grains are straight to crenulate. Zircon inclusions have been noted.

Feldspar (F) - Plagioclase is the dominant feldspar in the succession. It has an albite-oligoclase composition (confirmed by electron microprobe analysis) which shows no stratigraphic variation. Post-depositional albitisation of feldspars, however, is common and thus the values obtained may not reflect the original composition. K-feldspar is much less common. Crystals of perthite are ubiquitous, though rare. Feldspars are frequently altered to sericite and other clay minerals and may be replaced by calcite.

FORMATION	n	Q	F	L	Qm	F	Lt	Qp	Lvm	Lsm
Aberystwyth Grits	58	Mean Range	11.4 (3.29-19.62)	5.9 (0.62-15.1)	72.55 (54.35-89.44)	11.4 (3.29-19.63)	16.05 (4.84-31.43)	65.9 (6.66-90.9)	15.46 (0.0-60.0)	18.64 (2.66-43.75)
Allt Goch	1	Mean Range	9.83 (-)	2.89 (-)	78.03 (-)	9.83 (-)	12.14 (-)	76.19 (-)	4.76 (-)	19.05 (-)
Llangranog	10	Mean Range	9.34 (5.26-16.6)	3.59 (1.52-6.62)	77.33 (68.45-83.33)	9.34 (5.26-16.6)	13.33 (9.4-19.23)	73.1 (54.16-86.95)	11.42 (0.0-22.64)	15.48 (3.86-35.44)
Poppit Sands	51	Mean Range	7.11 (3.25-15.36)	7.67 (1.4-20.39)	72.42 (60.77-91.33)	7.11 (3.25-15.36)	20.47 (3.34-30.0)	63.59 (29.91-86.66)	13.89 (0.0-48.21)	22.52 (6.4-48.15)
Newport Sands	25	Mean Range	6.76 (2.2-11.46)	6.84 (2.81-19.39)	71.96 (52.38-85.8)	6.76 (2.2-11.46)	21.28 (14.38-39.62)	66.83 (43.8-86.59)	15.61 (1.6-50.6)	17.56 (0.0-36.36)
Parrog	1	Mean Range	7.75 (-)	5.43 (-)	82.17 (-)	7.75 (-)	10.08 (-)	46.15 (-)	23.08 (-)	30.77 (-)

(Q=total quartzose grains; F=total feldspar grains; L=total lithic fragments; Qm=monocrystalline quartz grains; Lt=total lithic fragments including polycrystalline quartzose grains; Qp=total polycrystalline quartzose grains; Lvm=total volcanic and metavolcanic lithic fragments; Lsm=total sedimentary and metasedimentary lithic fragments)

Table 5.1 Framework grain mode parameters of sandstones from the southern Welsh Basin.

Lithic Grains (L) - The most common lithic fragments are intraformational sedimentary rocks such as siltstone, mudstone and fine-grained sandstone. Volcanic rock fragments are commonly intermediate or acid in composition. Probable precursors of the felsic rock fragments were andesites, trachytes, granites and rhyolites. The rare metamorphic rock fragments are predominantly quartz and mica schists.

Accessory minerals (present in all formations) include muscovite, pyrite, chlorite-mica stacks, chlorite and zircon in decreasing order of abundance. Granophyre is present in all formations except for the Llangranog Formation. Other accessory minerals and their occurrences include tourmaline (Aberystwyth Grits Formation), biotite (Newport Sands Formation, Poppit Sands Formation, Aberystwyth Grits Formation), epidote (Poppit Sands Formation) and sphene (Newport Sands Formation, Aberystwyth Grits Formation).

### 5.5 Provenance and tectonic discrimination

The data are divided into two groups, the older formations which were deposited prior to the late Ashgill glacio-eustatic fall in sea level (Brenchley & Newall 1984) and the younger formations which were deposited during the low sea-level stand and the subsequent transgression. On the QFL diagram the older formations plot at the boundary of the continental block and recycled orogen provinces (Fig. 5.3). The younger formations also plot on this boundary but are skewed more towards the continental block province. According to Dickinson & Suczek (1979) sediments plotting within the continental block provenance are derived either from stable shields and platforms or from areas of uplift. Within recycled orogens, sediments are derived mainly from sedimentary strata and subordinate volcanics.

It is possible to increase the discrimination of the source area by assigning polycrystalline quartz (Qp) to the total lithics mode (Lt) in the Qm-F-Lt diagram (Fig. 5.4). The older formations plot in the recycled orogen province while

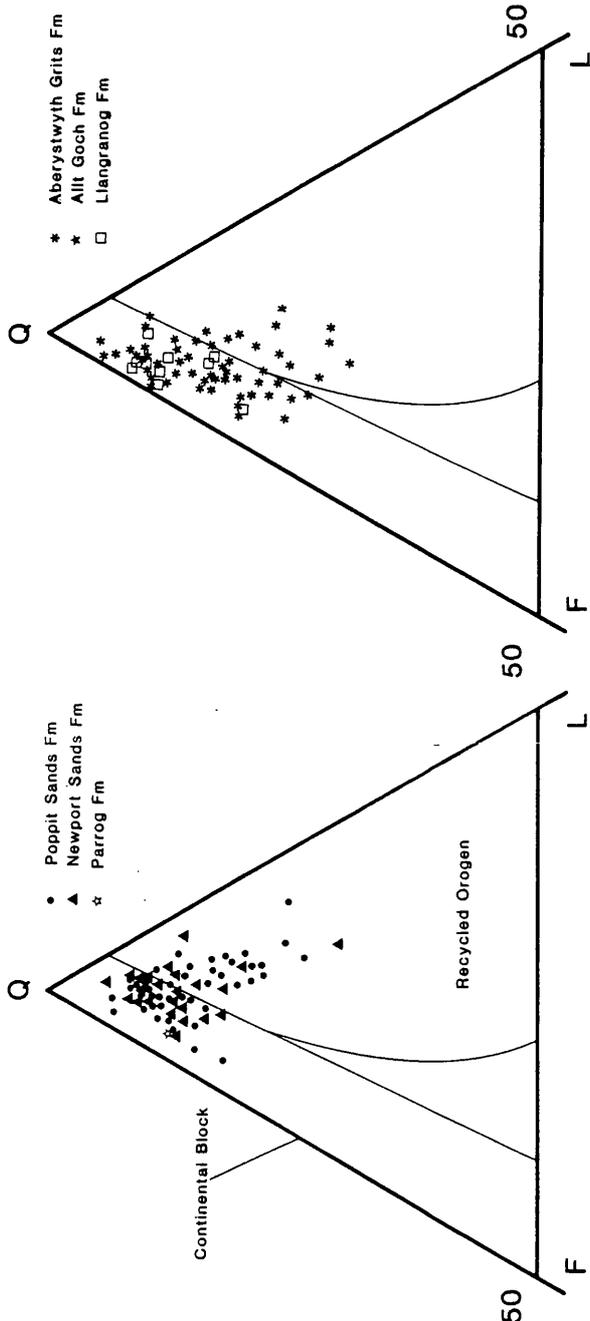


Figure 5.3 Ternary QFL plot showing mean framework modes for sandstones from the southern Welsh Basin: Q=total quartzose grains; F=total feldspar grains; L=total lithic fragments (after Dickinson & Suczek 1979).

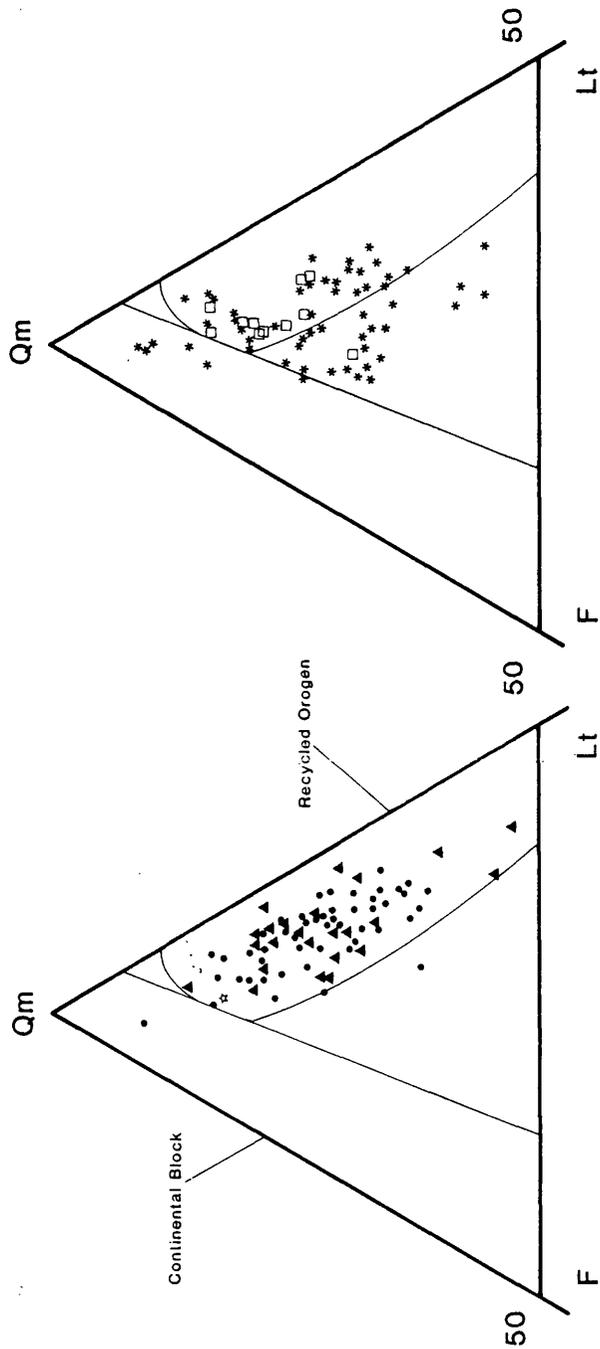


Figure 5.4 Ternary QmFt plot showing mean framework modes for sandstones from the southern Welsh Basin: Qm=monocrystalline quartz grains; F=total feldspar grains; Lt=total lithic fragments including polycrystalline quartzose grains (after Dickinson & Suczek 1979). Legend as for figure 3.

the younger formations plot on the boundary of the recycled orogen and continental block provinces. All of the data points plot in the quartzose recycled area and were probably derived from sediments whose ultimate source was cratonic. Sedimentological factors, however, may have locally enhanced quartz content and caution should therefore be exercised when interpreting the provenance of quartz-rich sediments (c.f. Mack 1984). Further discrimination of source is afforded by the Qp-Lvm-Lsm diagram showing the polycrystalline quartz (Qp) sub-population of the framework grain modes (Fig. 5.5). Almost all of the data points plot outside the two delimited fields indicating that the source area was neither a collisional nor an arc orogen.

#### 5.6 Microconglomerate Petrography

Twenty thin sections were examined by transmitted light microscopy and the electron microprobe, to determine the composition of selected lithic fragments. The majority of these thin sections were microconglomerates from either the Poppit Sands or Aberystwyth Grits formations. Intraformational sedimentary lithic clasts predominate although felsic fragments (typically rhyolites or andesites) and tuffs, trachytes and granitic clasts are also common. Both formations show similar assemblages of lithic fragments confirming the similar nature of the volcanic rocks in the source area and also its longevity.

An interesting feature is the occurrence in both rock suites of multicyclic sedimentary lithic fragments (cf. Zuffa 1987). This is in agreement with the evidence from analysis of the framework grains. The lack of mafic fragments is more problematic. Some of the volcanic centres in the source area contained extensive basalt sheets and pillow lavas (e.g. Fishguard Volcanic Complex) and yet there is very little evidence of this in the sandstone mineralogy of the area. This is either a function of the lower preservation potential of basaltic fragments, as opposed to that of fine-grained felsic fragments, or else that the basalts did not, at the

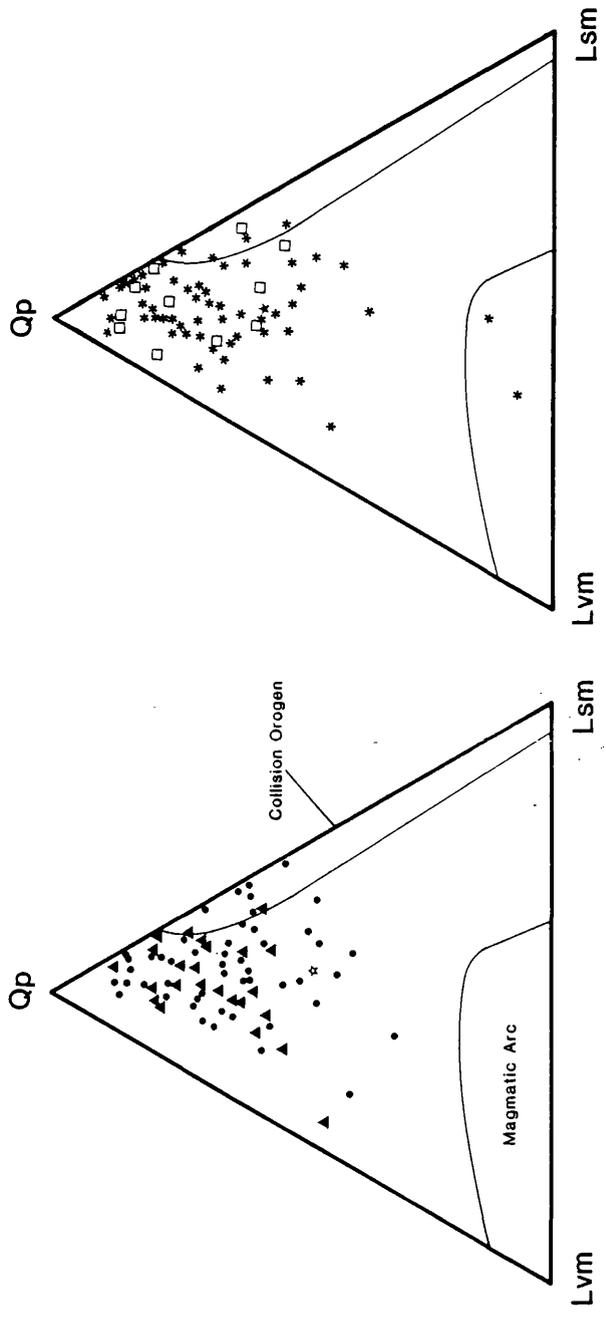


Figure 5.5 Ternary QpLvLsm plot showing mean framework modes for sandstones from the southern Welsh Basin: Qp=polycrystalline quartzose grains; Lvm=total volcanic and metavolcanic lithic fragments; Lsm=total sedimentary and metasedimentary lithic fragments (after Dickinson & Suczek 1979). Legend as for figure 3.

time of deposition, form part of the source area.

Three polished sections (one from the Poppit Sands Formation and two from the Aberystwyth Grits Formation) were examined using the electron microprobe to determine the chemical composition of some of the volcanic lithic clasts. A total of seven clasts were analysed and their total alkali and  $\text{SiO}_2$  contents determined. Based on these values, the lithic fragments may be classified as dacites, basalts and mugearites (cf. Cox et al. 1979). The samples suggesting a mugearite composition were both from the Aberystwyth Grits Formation and this is entirely in keeping with a possible derivation from the Skomer Volcanic Group (Lower Llandoverly) (Stillman & Francis 1979; Thorpe et al. 1989). Basalts and dacites are recorded from the majority of the Tremadoc-Llandeilo volcanic episodes (Allen 1982)

### 5.7 Chemical Classification of Rocks

The major and trace element geochemistries for twenty-nine mudstones and thirty two sandstones were determined by means of X-ray fluorescence using a Phillips PW1400 X-ray generator following the procedure of Marsh et al. (1983) and Weaver et al. (1983). The sandstones chosen were all medium-grained. These results were then used to provide the basis for geochemical analysis of the sediments.

The majority of the sandstones have a  $\text{SiO}_2$  range of 70.7–85.67 wt%, low  $\text{Fe}_2\text{O}_3$  (total Fe as  $\text{Fe}_2\text{O}_3$ ) and MgO contents of between 1.9 and 3.01 (Table 5.2). Chemical classification of the sandstones indicates that they are quartz-rich to quartz-intermediate sandstones (Fig. 5.6). Large-ion-lithophile (LIL) elements such as K, Rb, Sr and Th show a range of abundances.

Variations in the major element geochemistry of both the sandstones and mudstones are shown on Harker diagrams (Fig. 5.7). The data point distributions of the mudstones tend to cluster while the sandstones have more linear distributions.

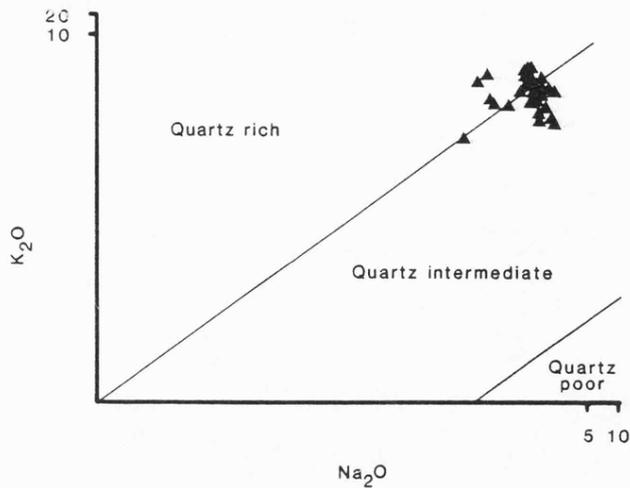


Figure 5.6 Analysis of quartz-richness of southern Welsh Basin sandstones based on major element geochemistry (after Crook 1974).

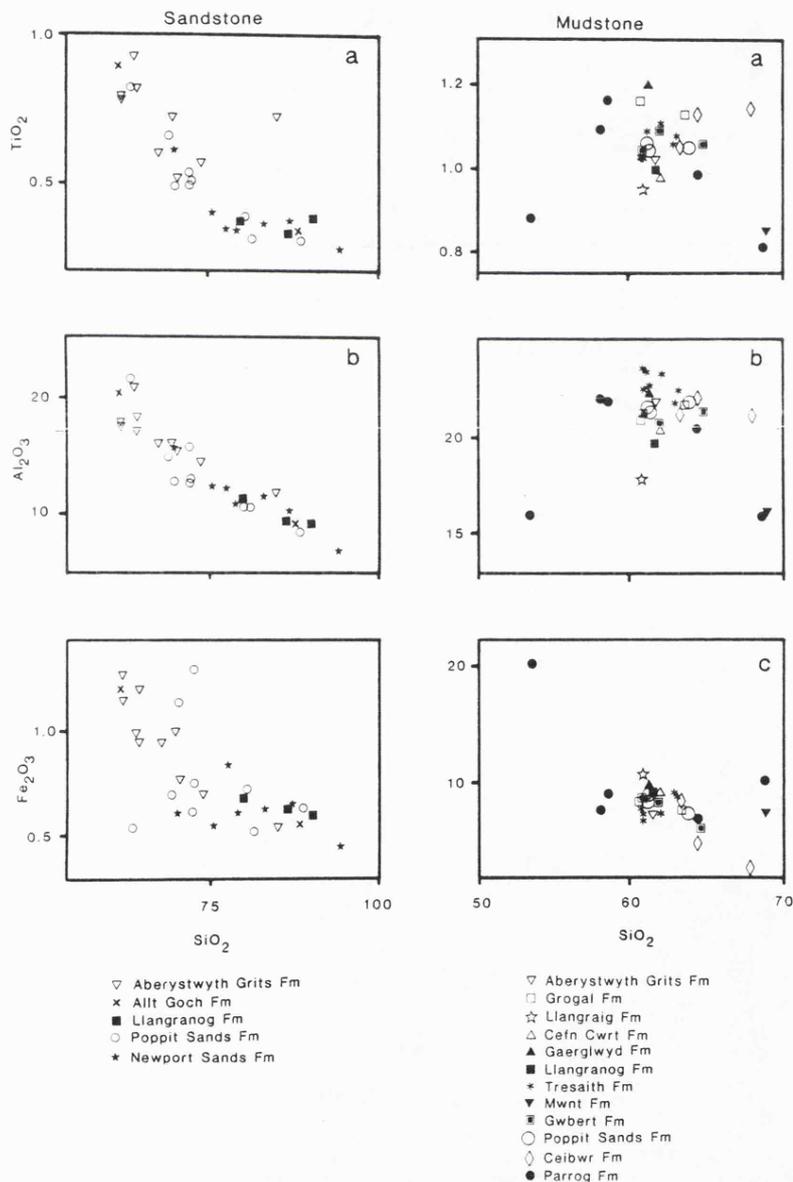


Figure 5.7 (a-c) Harker variation diagrams of major elements for sandstones and mudstones from the southern Welsh Basin.

(a) SANDSTONES

Pellet No.	L11504	L11008	L11009	L10981	L11593	L11585	L11581	L11587	L11588	L11595
Formation	N.S.	N.S.	P.S.	P.S.	Llan.	Llan.	A.Goch	A.Grits	A.Grits	A. Grits
SiO <sub>2</sub>	78.0	70.5	73.0	80.7	86.7	80.1	62.4	70.3	85.3	65.1
TiO <sub>2</sub>	0.34	0.62	0.51	0.39	0.33	0.38	0.91	0.73	0.74	0.84
Al <sub>2</sub> O <sub>3</sub>	12.2	15.6	13.0	10.6	9.4	11.3	20.4	16.0	11.9	18.3
Fe <sub>2</sub> O <sub>3</sub>	0.8	0.6	0.7	0.7	0.6	0.7	1.2	1.0	0.5	1.2
MnO	0.04	0.04	0.04	0.05	0.1	0.04	0.08	0.15	0.11	0.03
MgO	1.9	1.8	2.2	2.0	1.1	1.6	2.7	2.4	1.1	2.6
CaO	0.1	0.3	0.3	0.2	0.2	0.2	0.4	1.2	0.1	0.2
Na <sub>2</sub> O	0.3	1.3	1.6	1.4	2.3	2.0	1.3	1.5	1.1	1.5
K <sub>2</sub> O	0.39	2.59	1.35	1.2	0.7	1.01	3.18	2.11	1.68	2.25
P <sub>2</sub> O <sub>5</sub>	0.13	0.38	0.23	0.14	0.1	0.14	0.28	0.16	0.09	0.15
Total	94.25	93.78	92.92	97.39	101.57	97.38	92.73	95.48	102.63	92.27
V	98	98	66	64	39	46	130	91	69	105
Cr	59	59	73	48	18	27	85	68	69	68
Ba	406	406	225	208	104	142	434	269	298	281
La	21	21	15	17	BDL	BDL	BDL	BDL	BDL	BDL
Ce	47	47	32	38	BDL	BDL	BDL	BDL	BDL	BDL
Nd	22	22	16	17	BDL	BDL	BDL	BDL	BDL	BDL
Nb	8	8	8	7	5	7	15	12	12	15
Zr	140	140	145	92	110	124	219	205	517	296
Y	28	28	23	23	16	14	40	28	23	32
Sc	63	63	51	55	37	40	87	70	45	60
Rb	72	72	46	39	29	41	132	92	72	88
Th	10	10	8	5	4	7	14	10	8	11
Ga	14	14	13	10	10	9	24	20	13	20
Zn	47	47	74	61	52	56	114	83	63	85
Ni	16	16	19	15	11	17	39	36	20	47

(N.S.=Newport Sands; P.S.=Poppit Sands; Llan=Llangranog; A. Goch=Allt Goch; A. Grits=Aberystwyth Grits; BDL=Below Detection Limits; Total Fe as Fe<sub>2</sub>O<sub>3</sub>)

Table 5.2 Representative chemical analyses of sandstones and mudstones from the southern Welsh Basin. Major oxide in wt %, trace elements in ppm. Full chemical analyses are in Appendix 1.

## (b) MUDSTONES

Pellet No.	L11002	L10974	L10994	L10989	L10995	L10979	L10997	L11004	L11000	L10977
Formation	Parrog	Coibur	P.S.	Gubert	Tresaith	Tresaith	Georglywd	Cefn Curt	Llangraig	Grogal
SiO <sub>2</sub>	58.1	63.3	61.1	61.9	62.0	61.0	61.2	62.0	60.8	60.7
TiO <sub>2</sub>	1.09	1.05	1.06	1.09	1.11	1.03	1.2	0.98	0.95	1.16
Al <sub>2</sub> O <sub>3</sub>	22.0	21.3	21.6	20.8	23.4	23.6	22.5	20.6	17.9	21.0
Fe <sub>2</sub> O <sub>3</sub>	7.7	8.4	8.2	8.3	7.3	7.2	9.5	10.7	9.0	8.3
MnO	0.07	0.07	0.08	0.05	0.09	0.03	0.06	0.18	0.2	0.15
MgO	2.5	2.0	2.1	2.2	2.0	1.8	1.8	2.4	2.7	1.8
CaO	0.2	0.1	0.2	0.2	0.1	0.1	0.2	0.2	0.1	1.1
Na <sub>2</sub> O	3.1	1.3	1.4	1.3	1.6	1.6	1.0	1.5	1.6	1.4
K <sub>2</sub> O	4.38	3.48	3.7	3.4	3.75	3.78	3.93	3.4	2.47	3.76
P <sub>2</sub> O <sub>5</sub>	0.12	0.10	0.21	0.11	0.07	0.07	0.24	0.17	0.18	0.53
Total	99.35	101.18	99.68	99.27	101.61	100.18	101.63	100.31	97.62	99.86
V	170	163	186	142	178	162	164	135	113	156
Cr	115	99	103	171	122	116	172	82	77	112
Ba	575	487	545	447	558	658	655	437	291	560
La	36	24	30	32	38	42	47	36	23	41
Ce	68	45	64	67	80	82	100	78	49	93
Nd	29	19	27	27	29	33	43	33	21	46
Nb	19	18	17	18	19	16	23	17	16	21
Zr	237	223	206	279	183	168	217	182	242	246
Y	36	35	35	31	30	31	39	33	24	47
Str	117	105	108	74	119	125	116	89	74	116
Rb	166	148	151	131	192	187	158	137	92	146
Th	16	15	11	14	12	14	17	14	10	15
Ga	28	26	28	22	32	28	37	21	24	26
Zn	113	108	111	98	105	112	126	115	127	99
Ni	42	39	45	39	41	74	44	37	48	42

(N.S.=Newport Sands; P.S.=Poppit Sands; Llan=Llangranog; A. Goch=Allt Goch; A.Grits=Aberystwyth Grits;  
BDL=Below Detection Limits; Total Fe as Fe<sub>2</sub>O<sub>3</sub>)

This is a consequence of variability in grain-size within the sandstones and also their higher quartz contents relative to the mudstones. Older sandstone-dominated formations (e.g. Newport Sands Formation, Poppit Sands Formation) tend to be more quartz-rich (as noted earlier in the discussion of framework grains) and, therefore, they have less  $\text{TiO}_2$ ,  $\text{Al}_2\text{O}_3$ ,  $\text{Fe}_2\text{O}_3$  and  $\text{MgO}$  than younger formations (Fig. 5.7a,b,c,f). Certain elements, for example  $\text{MnO}$ ,  $\text{CaO}$  and  $\text{Na}_2\text{O}$ , have very poorly defined distribution patterns in the sandstones whereas in the mudstones their distribution is fairly constant (Fig. 5.7d,e,g). The  $\text{MnO}$  plot for the mudstones tends to have a vertical distribution pattern reflecting the variable concentrations of the element within the mudstones (Fig. 5.7d). There is no stratigraphic basis to the distribution. The mudstone-dominated formations are significantly enriched in  $\text{K}_2\text{O}$  (Fig. 5.7h). This is related to their increased phyllosilicate content as seen in the dominance of chlorite-mica stacks. The plots of both  $\text{TiO}_2$  and  $\text{Al}_2\text{O}_3$  both show even distributions of those elements with respect to  $\text{SiO}_2$  for the sandstone samples (Fig. 5.7a,b). This is probably a function of sediment maturity. The single point (i.e. Aberystwyth Grits Formation) on the  $\text{TiO}_2$  plot which is not on this curve is possibly rich in rutile (Fig. 5.7a). The  $\text{Fe}_2\text{O}_3$  sandstone plot shows a similar distributional pattern to the  $\text{TiO}_2$  and  $\text{Al}_2\text{O}_3$  plots (Fig. 5.7c). The displacement, however, of three points from the Poppit Sands Formation may be a result of iron enrichment in some of the samples. Thin-section analysis of sandstones from this formation has revealed that some clay and siltstone lithic fragments may be haematized. These same data points are also displaced in the  $\text{MgO}$  plot although the reason for this is not known. In the  $\text{CaO}$  plot the samples from the Aberystwyth Grits Formation tend to cluster away from the rest of the data points (Fig. 5.7e). In thin section this formation contains a patchy calcite cement and this is the probable cause.

Samples from the oldest sandstone formation (i.e. Newport Sands Formation) tend to have lower levels of  $\text{Na}_2\text{O}$  than the bulk of the samples (Fig. 5.7g,h). The same points also tend

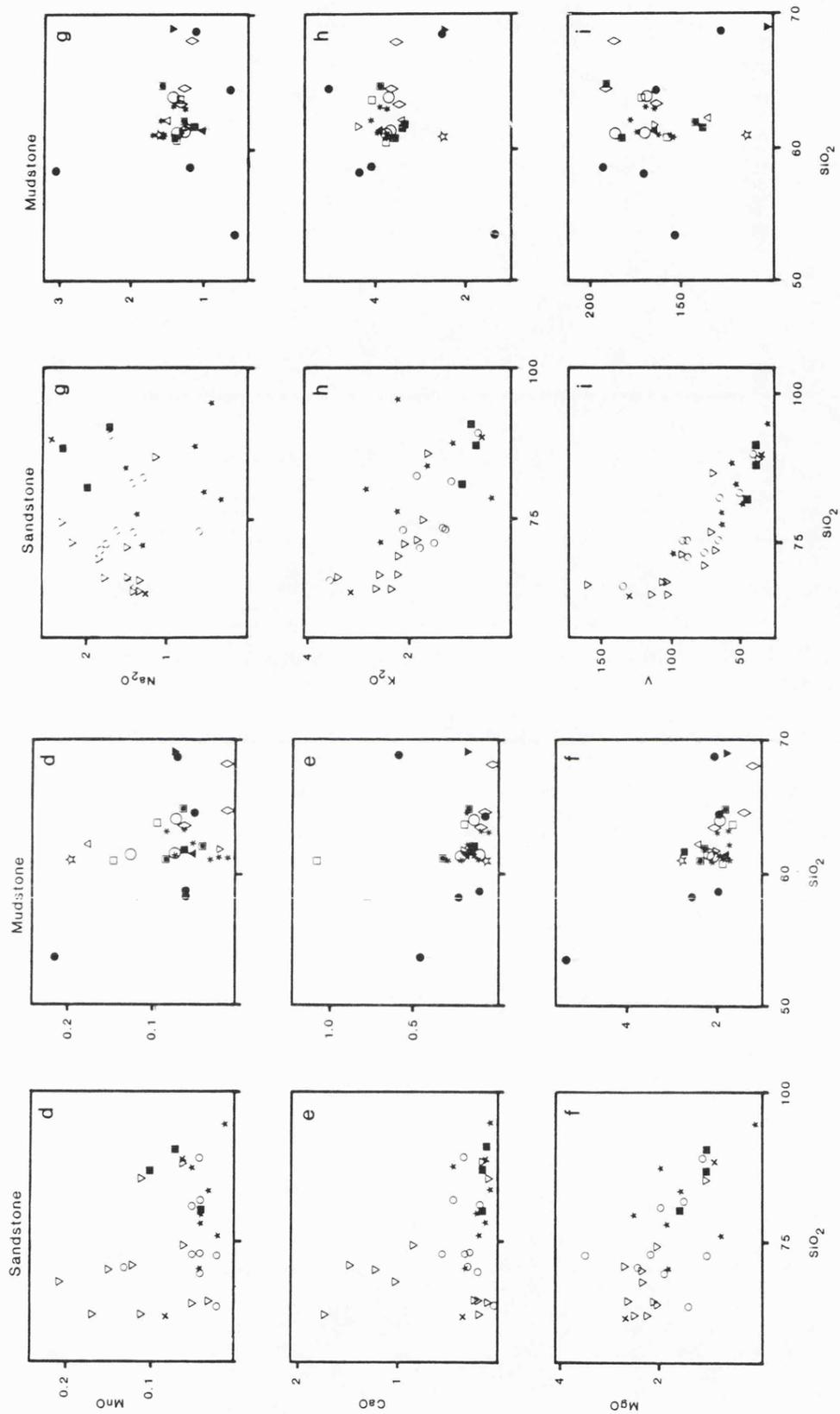


Figure 5.7 (d-i) Harker variation diagrams of major elements for sandstones and mudstones from the southern Welsh Basin.

to plot outside of the main  $K_2O$  trend. Both of these features may be related to the age of the formation and the fact that it was deposited soon after the cessation of volcanic activity in the source area. The K-rich and Na-poor nature of the samples is a reflection of the possible higher quantities of alkali feldspar in the sandstones.

Both V and  $TiO_2$  were plotted against  $SiO_2$  for sandstones and mudstones (Fig. 5.7a,i). In the plots V and  $TiO_2$  show similar patterns suggesting that magnetite was their precursor phase. Plotting V against  $TiO_2$  shows a high correlation particularly for the sandstones (Fig. 5.7j). A real difference, and poorer correlation, would be expected if the parent phase was rutile, clinopyroxene or garnet.

Plotting Ni and Cr against  $SiO_2$  for the sandstones and mudstones reveals that both show similar distributions to the  $Fe_2O_3$  and MgO plots (Fig 5.7k,l). All of these elements (Cr, Ni, Mg and Fe) occur in basic igneous rocks and the cohesion of the plots suggests that they were derived from such a source. Minimal levels of weathering and alteration are also suggested by the cohesion of the plots; the MgO levels have not been altered too much by clay mineral chemistry nor have the  $Fe_2O_3$  levels been affected by oxides. Thus both time and chemical weathering processes are constrained suggesting either a proximal relationship to source or else relatively little geochemical alteration in the depositional environment.

### 5.8 Geochemical analysis of provenance

Geochemical analysis of sediments may indicate the plate tectonic setting. Three main tectonic provenances are defined by Roser & Korsch (1986): (a) Passive Continental Margin (PM) - Mineralogically mature (quartz-rich) sediments deposited in plate interiors at stable continental margins or intracratonic basins (equivalent to the "trailing-edge tectonic setting" of Maynard *et al.*, 1982); (b) Active Continental Margin (ACM) - Quartz-intermediate sediments

derived from tectonically active continental margins on or adjacent to active plate boundaries (e.g. trench, forearc and backarc settings); (c) Oceanic Island Arc (OIA) - Quartz-poor volcanogenic sediments derived from oceanic island arcs (i.e. sediments derived from an island arc source and deposited in a variety of settings including forearc, intra-arc and backarc basins and trenches).

The following sections consider the geochemistry of the Welsh Basin sediments in terms of provenance utilising a variety of different approaches.

### 5.8.1 Quartz richness

Crook (1974) subdivided sandstones on the basis of  $\text{SiO}_2$  content and the relative  $\text{K}_2\text{O}/\text{Na}_2\text{O}$  ratio into three classes and assigned each to a plate tectonic environment. All of the sandstone samples from the study area may be classified as either quartz-rich (average 89%  $\text{SiO}_2$ ,  $\text{K}_2\text{O}/\text{Na}_2\text{O} > 1$ ) or quartz-intermediate (average 68-74%  $\text{SiO}_2$ ,  $\text{K}_2\text{O}/\text{Na}_2\text{O} < 1$ ) (Table 5.3). Based on the  $\text{K}_2\text{O}/\text{Na}_2\text{O}$  ratio, four of the sandstone formations may be classified as quartz-rich. The  $\text{SiO}_2$  wt.% of these samples are, however, lower than the average value of 89% suggested by Crook (1974) and are closer to the values for quartz-intermediate sandstones. It is best, therefore, to consider the samples as falling on the quartz-rich/quartz-intermediate boundary. Modern equivalents of quartz-rich sediments adjoin Atlantic-type continental margins on the trailing edge (PM) of continents, whereas quartz-intermediate sediments are more indicative of Andean-type (ACM) margins.

### 5.8.2 Major element analysis of sandstones (Fig. 5.8)

Bhatia (1983), in a study of the geochemistry of sandstones from Australia, devised a series of plots to differentiate four main tectonic settings. The most discriminating parameters are  $\text{TiO}_2$  wt%,  $\text{Al}_2\text{O}_3/\text{SiO}_2$ ,  $\text{K}_2\text{O}/\text{Na}_2\text{O}$  and

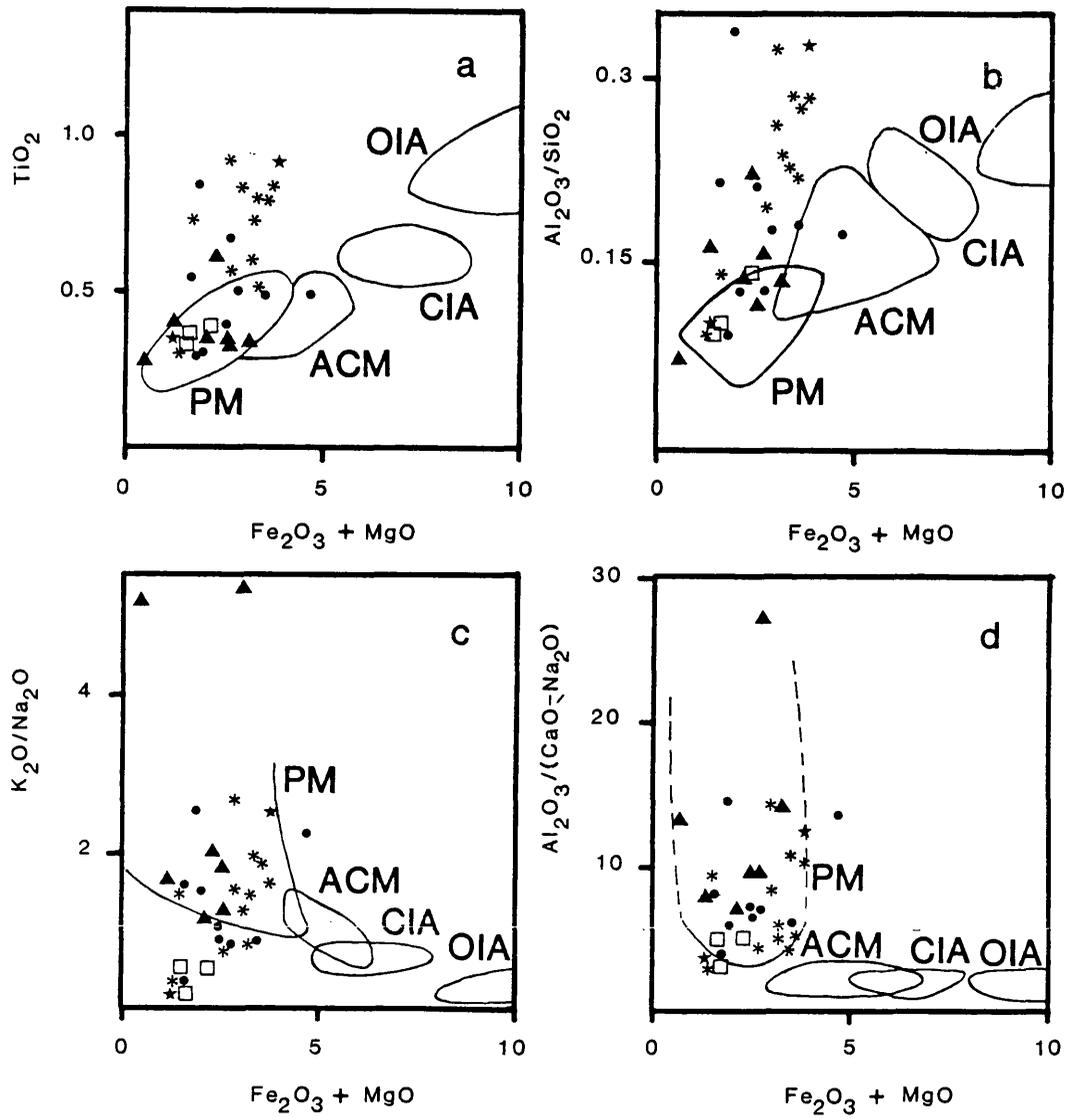


Figure 5.8 Bivariate plots for the discrimination of plate tectonic setting of sandstones from the southern Welsh Basin: PM=passive margin; ACM=active continental margin; CIA=continental island arc; OIA=oceanic island arc (after Bhatia 1983). Legend as for figure 3.

Formation	n	SiO <sub>2</sub> wt%	K <sub>2</sub> O/Na <sub>2</sub> O
Aberystwyth Grits	12	70.7	1.26
Allt Goch	2	75.35	1.03
Llangranog	3	85.67	0.43
Poppit Sands	9	74.9	1.19
Newport Sands	7	81.21	2.19

Table 5.3 Mean SiO<sub>2</sub> wt% and K<sub>2</sub>O/Na<sub>2</sub>O values of sandstone-dominated formations from the southern Welsh Basin (values after Crook 1974).

$\text{Al}_2\text{O}_3/(\text{CaO}+\text{Na}_2\text{O})$  ratios all plotted against  $\text{Fe}_2\text{O}_3+\text{MgO}$  wt%. In the  $\text{TiO}_2$  wt% and  $\text{Al}_2\text{O}_3/\text{SiO}_2$  plots (Fig. 5.8a, b) the majority of the points fall within the passive margin (PM) field although some also plot within the active continental margin field (ACM). The observed vertical distribution of the points is interpreted as a function of interelement variations.

Plotting the ratio of  $\text{K}_2\text{O}/\text{Na}_2\text{O}$  produces a distribution where most of the formations fall within the PM field (Fig. 5.8c). Displacement is here affected by the degree of maturity of the sediments, maturity being directly reflected in the relative feldspar ratios. In Fig. 5.8d the majority of the formations plot within the PM field. Again, displacement of data points is a function of sediment maturity.

The distribution of most of the points in or around the PM field of Bhatia (1983) suggests that they may be derived from Atlantic-type rifted continental margins, remnant ocean basins adjacent to collision orogens and inactive or extinct convergent margins. Within the PM field sediments are generally highly matured, being derived from the recycling of older sedimentary and metamorphic rocks on platforms or recycled orogens.

### 5.8.3 K/Rb diagram (Fig. 5.9)

This plot may be used to distinguish those sediments derived from rocks of acid and intermediate compositions from those derived from rocks of basic composition. The relatively high K/Rb of the sediments from the Welsh Basin is indicative of derivation from acid and intermediate source rocks with some input from basic sources.

### 5.8.4 $\text{K}_2\text{O}/\text{Na}_2\text{O}$ vs $\text{SiO}_2$ Diagram (Fig. 5.10)

The mudstone samples from the Welsh Basin plot astride the Active Continental Margin (ACM) and the Passive Margin (PM)

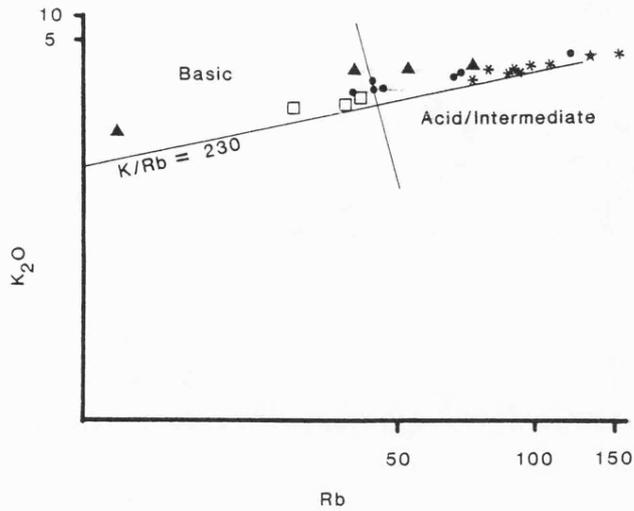


Figure 5.9 Distribution of K (log wt%) and Rb (ppm) in the southern Welsh Basin sandstones relative to a K/Rb ratio of 230 (= Main Trend of Shaw 1968). Boundary line between acid/intermediate and basic compositions after Floyd & Leveridge (1987). Legend as for figure 3.

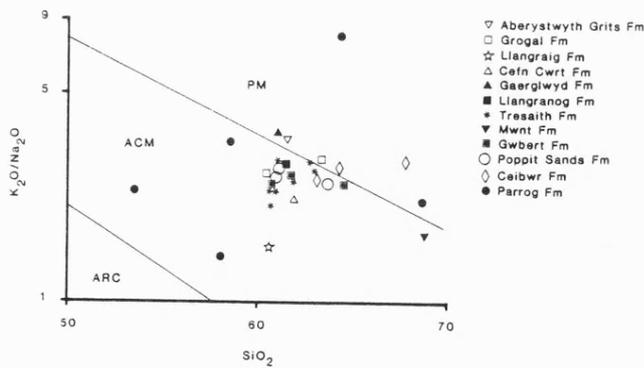


Figure 5.10 Tectonic discrimination diagram for mudstones from the southern Welsh Basin: PM=passive margin; ACM=active continental margin; ARC=arc (after Roser & Korsch 1986).

border of Roser & Korsch (1986). This position suggests that the presence of arc-derived material may be discounted. It should be noted, however, that the distribution of the data points does not show any stratigraphic trend. According to Roser & Korsch (1986) the location of data points on the diagram is primarily controlled by the nature of volcanism, the extent of plutonism and related erosional levels. The effect of mineralogical maturation through sediment recycling is a secondary consequence. The data points from the Welsh Basin correspond with the Greenland and Torlesse terranes (New Zealand) of Roser & Korsch (1986). The Greenland Terrane is a recycled quartzose sandstone (PM) with most of the data points lying within the PM field. The Torlesse Terrane, however, was derived from an ACM tectonic setting compatible with the quartz- intermediate nature of the sandstones.

#### 5.8.5 Trace element concentrations

Very high levels of Cr (e.g. 100-1500ppm) and Ni (e.g. 50-600) have been used by a variety of authors (e.g. Hiscott 1984; Haughton 1988; Wrafter & Graham 1989) to indicate an ultramafic provenance for the sediments. The low levels of Cr (17-125 ppm) and Ni (3-47 ppm) recorded in the Welsh sandstone- dominated formations suggests either some basic input into the system or else that the trace elements could have travelled into the depositional basin as adsorbed ions on clays. Vanadium levels are relatively high in both the sandstones (31-159 ppm) and mudstones (102-193 ppm). These levels are higher than the levels commonly recorded in sandstones (20 ppm) and given that V is concentrated in basic rocks they suggest some basic input into the depositional system.

Bhatia (1985) used trace elements to geochemically determine the tectonic setting of mudstones. He distinguished four main tectonic provenances using concentrations and interelement ratios of Nb, Zr, Y, Rb, Sr, Th, Ba, Cr and Ni. The results obtained herein are not definitive with all four

Element	Mean (n=29)	OIA	CIA	ACM	PM
Nb	17.27	3.7	9.0	16.5	15.8
Zr/Nb	12.2	38.0	21.0	11.0	10.0
Nb/Y	0.51	0.17	0.35	0.5	0.54
Rb/Sr	1.47	0.29	1.31	2.9	5.8
Th	13.34	5.5	16.2	28.0	22.0
Zr/Th	15.82	28.0	12.0	7.0	7.0
Ba/Sr	6.33	2.5	6.3	8.7	17.6
Cr	111.38	39.0	55.0	58.0	100.0
Ni	42.24	15.0	18.0	26.0	36.0

Table 5.4 Trace element geochemical parameters for mudstone-dominated formations from the southern Welsh Basin: OIA - Oceanic Island Arc; CIA - Continental Island Arc; ACM - Active Continental Margin; PM - Passive Margin; all values after Bhatia 1985)

tectonic settings being indicated. The high Ni and Cr contents suggest a passive margin origin while the Rb/Sr value is closest to that of a continental island arc (Table 5.4). The Th and Zr/Th values fall between those suggesting a continental island arc and an oceanic island arc, but are closer to the former. The Ba/Sr ratio suggests a continental island arc setting. The remaining three factors all compare most favourably with the active continental margin tectonic setting (Table 5.4).

#### 5.8.6 Multi-element diagram normalised to Post Archean Shale (Fig. 5.11)

Multi-element diagrams may be used to examine the distribution of trace elements in mudstones. The plot compares a range of elements from seven of the mudstone-dominated formations against a normalised post-Archean average shale (PAAS; Taylor & McLennan 1985). The elements are arranged such that those elements mainly derived from acidic source rocks plot on the left-hand side of the diagram while those derived from basic and ultrabasic source rocks are plotted on the right-hand side. The main feature to note is the degree of conformity of the values with those of the PAAS. Most values are evenly spread out around the PAAS although certain elements (e.g. V, Cr, Fe) do suggest a greater amount of basic input.

### 5.9 Discussion

It has been suggested that the tectonic setting of the Welsh Basin was a passive margin prior to the Tremadoc, an active margin back-arc basin from the early Ordovician to the Caradoc and a non-volcanic active margin or collision zone basin from the Ashgill to the Devonian (Pickering *et al.* 1988; Woodcock *in press*). Thus the tectonic setting of the basin for the entire period of deposition of the sediments was that of an active margin. Volcanic activity was particularly pronounced in the early Ordovician but had

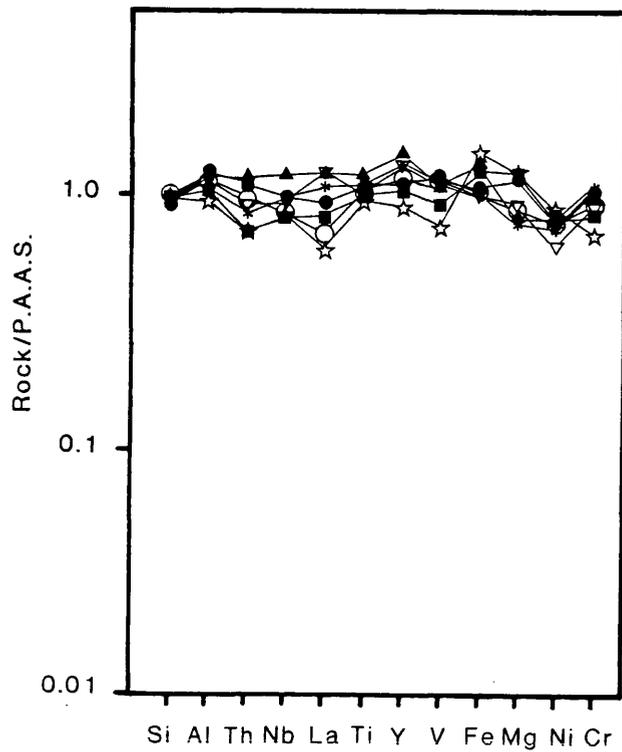


Figure 5.11 Multi-element diagram normalised to Post-Archean average shale (after Taylor & McLennan 1985). Legend as for figure 10.

ceased, in the southern Welsh Basin, by the Llanvirn. The eruption of the Skomer Volcanic Group (early Llandovery) was the only volcanic episode coeval with deposition within the southern Welsh Basin. Caradoc volcanism was confined to North Wales and was not a detrital source for the sediments of the depositional area.

The signatures from the sedimentary provenance indicators, however, do not agree with the active margin setting. Both the framework grains and the major element geochemistry of the sandstones suggest a passive margin tectonic setting for the area. The mudstone geochemistry, particularly that of the trace elements, is more variable, suggesting a variety of tectonic settings ranging from continental island arcs to passive margin. What is remarkable about both sets of data, however, is the lack of stratigraphic variation between the data sets. Indeed, the provenance signature of the sediments shows a remarkable degree of uniformity over the entire period of deposition within the depositional basin (c. 35 Ma).

As mentioned earlier, palaeocurrent evidence within the region shows that sediment was chiefly derived from the south and southeast. Transport directions are similar for the majority of the formations; the only exceptions being some south-directed current directions to the north of Aberystwyth (McCann & Pickering 1989) and westerly-directed currents in some of the mudstone-dominated formations, these latter produced as a result of lateral transport of sediment at lobe margins (see Chapter 4).

The passive margin signature was derived, therefore, from this southern landmass. There are two ways in which this could have occurred. Firstly, detritus may have been derived from the calc-alkaline Precambrian basement and associated Lower Palaeozoic rocks and therefore records the original tectonic signature (PM) of this basement and not the setting of the active basin. Alternatively, the sediment could have been derived distally from the trailing passive margin of the southern landmass. The landmass is generally considered to

have been narrow, separating the Welsh Basin from the Rheic Ocean to the south. During the period of deposition there was an active continental margin to the north of the Welsh Basin (with the closing Iapetus Ocean) while to the south of the microcontinent of Eastern Avalonia there was a passive margin to the Rheic Ocean. This situation is somewhat analogous to the southern end of present day South America where the Chilean side is a leading margin (ACM) while the Argentinian side is a trailing margin (PM) (Fig. 5.12). It is also interesting to note that both the southern tip of South America and the microcontinent of Eastern Avalonia (which contains the Welsh Basin) are of similar size (Fig. 5.12). In such situations, where two disparate settings exist adjacent to one another, then the tectonic setting, as deduced from beach sands, may be an amalgamation of both (Potter 1984, 1986). The analogy is supported by evidence suggesting that sediment was carried across the Tornquist-Teisseyre Lineament from the area of Baltica into the Welsh Basin (Pickering 1989). It is, however, difficult to determine with any degree of certainty which of the two models are the most likely. However, it does seem unlikely that sediment from the passive margin could find its way into the back-arc Welsh Basin and thus it is more probable that the reflected signature is a relict one.

#### 5.9.1 Sandstone provenance

Much of the sand-grade sediment was deposited as part of elongate turbidite systems developed from point sources in the south/southeast. As noted earlier such elongate turbiditic bodies tend to derive their sediment from point sources, for example, large rivers or delta systems. Turbidite deposition is very much controlled by eustatic changes in sea level (Mutti & Normark 1987). Maximum input of detritus is during low sea-level stands when sediment sources (e.g. rivers) can prograde over the shelf area and directly funnel sediment loads into the deeper marine basins (Stow et al. 1985; Vail et al. 1977). This sediment discharge, therefore, would be a more accurate reflection of

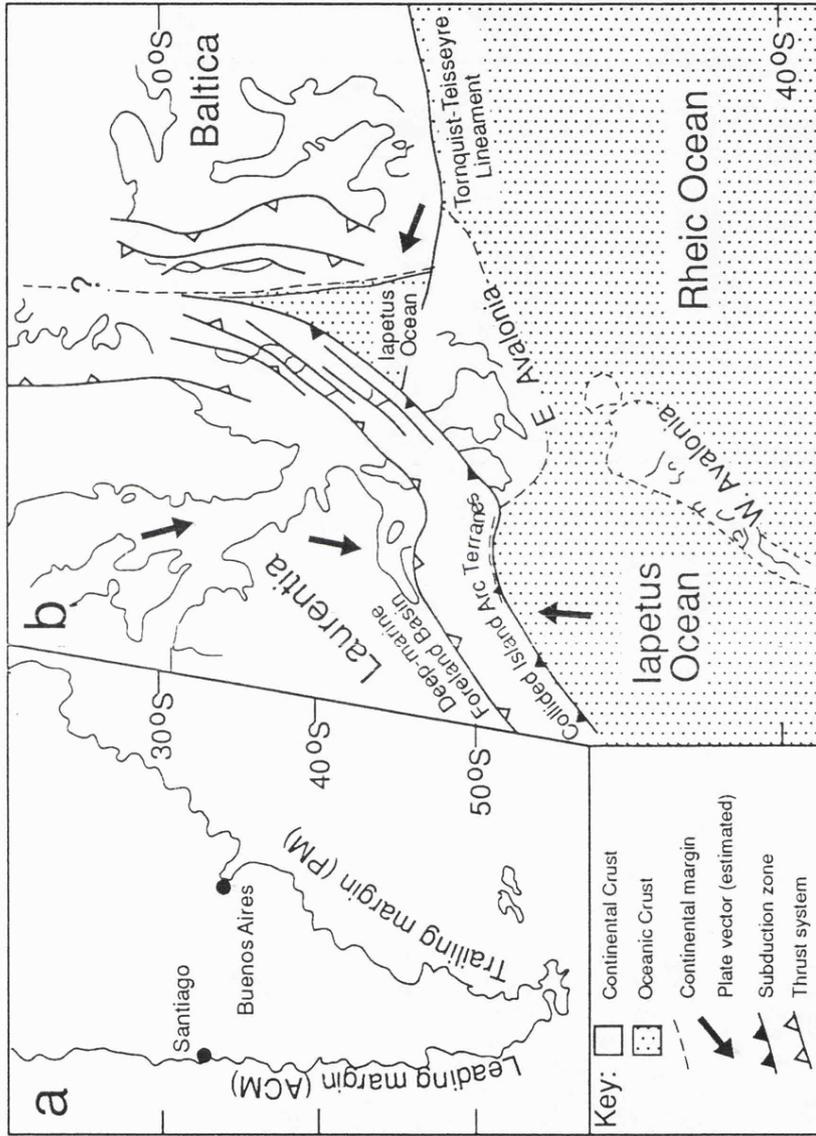


Figure 5.12 Comparative diagram showing (a) present day tectonic setting of South America, and (b) Ashgill-Llandovery palaeogeographic reconstruction of the North Atlantic region ((b) modified from Pickering 1989 and Vannier et al. 1989). The key refers to (b) only.

the tectonic setting suggested by the hinterland rather than that of the depositional basin.

A qualitative approach, based on individual lithic fragments, does provide more information about the volcanic successions. Unfortunately, given that many of the volcanic successions are lithologically similar, it is not possible to trace the erosion of particular volcanic centres. Furthermore, while Zuffa (1987) has subdivided volcanic fragments into coeval and ancient types, this classification has only been used for modern successions. It is doubtful, given the significant degree of alteration which can occur in volcanic fragments, whether this classification could be applied to ancient sequences. Certainly in this situation, with the majority of the volcanic fragments being classified as "felsic", it is probable that they could have been derived from Precambrian as well as Ordovician sources.

The limited sample size of Bhatia's (1983) sandstone provenance model (69 samples used) may explain the relatively poor correlation obtained when using the model to examine the sediments of the southern Welsh Basin. While all 32 Welsh samples plotted in or around the passive margin field there tended to be a vertical spread for which there was no explanation. Although this was primarily a function of inter-element variability and/or maturity neither reason was taken into consideration by the fields of Bhatia (1983), nor was there any explanation in the text that such a spread might occur. It is suggested that caution be exercised when applying this model to ancient sequences, about which little is known in terms of provenance, as it may be misleading.

### 5.9.2 Mudstone provenance

The depositional environments of the mudstones are more variable than those of the sandstones, occurring both as turbidite-related deposits and also as general background sedimentation (e.g. hemipelagites) within the basin. Some of the sediment may also be redeposited sediment derived from

shelf areas and possibly input as a result of storm activity on the shelf. The more disparate tectonic settings, therefore, may be related to the fact that the mudstones are derived, not only from point sources (in the case of muds which formed part of the turbidite systems) but also from the more general area of the shelf.

None of this, however, explains the diverse provenance attributions derived from the use of the trace element geochemistry parameters of Bhatia (1985). The spread of the results suggests that there are some fundamental flaws in Bhatia's (1985) arguments.

According to Bhatia (1985), the active continental margin (ACM) and passive margin (PM) mudstones are similar in most immobile trace elements and may be distinguished from the mudstones of other tectonic settings by their significantly higher Th and Nb percentages and Nb/Y ratio and lower Zr/Th ratio. They may be distinguished from each other by the higher Rb/Sr and Ba/Sr ratios and Cr and Ni percentages of the passive margin setting. The increased Cr and Ni in passive margin tectonic settings is a result of enrichment and adsorption of these elements with the increased phyllosilicate content. The decrease in Rb/Sr and Ba/Sr is due to the loss of Sr and feldspar with the increased weathering and recycling of passive margin type sediments.

The geochemical behaviour, however, of some of these elements is extremely variable. For example, Dimberline (1987), notes that Rb/Sr ratios may be reduced due to the presence of additional Sr incorporated in diagenetic carbonate. The presence of diagenetic carbonates in some of the mudstone-dominated formations may, therefore, have depressed the values. As noted earlier, the distribution of Sr is affected to an extent by the presence of Ca. Fairbridge (1972) also notes that the Sr content of sedimentary rocks is variable because of the many influences on Sr in low temperature deposition. Mudstones seem to have an ability to concentrate Sr due to ion exchange properties of the clay minerals and concentrations of up to 298 ppm have been

reported (Fairbridge 1972). This would cast considerable doubt on the advisability of using Sr, either alone or in combination with other elements, for provenance determination in mudstones. The geochemistry of Ba is very close to that of Sr and, therefore, it too would appear to be an unreliable provenance indicator.

The behaviour of Rb is largely controlled, in sedimentary processes, by its adsorption on clay minerals (Fairbridge 1972). While both the degree of adsorption and the presence or absence of certain clay minerals may be related to provenance, the link is not a firm one and, therefore, it does not appear to be a particularly reliable element to use.

Dimberline (1987) noted that high Ni contents could be produced by having Ni concentrated preferentially in organic matter. The presence of organic matter in laminated hemipelagites in the Gaerglwyd Formation and parts of the Parrog and Tresaith formations could, therefore, have resulted in a high Ni value.

Based on geochemical criteria alone, the model of Bhatia (1985) is questionable. In the present study area 29 samples, were used while Bhatia (1985) used a total of 23 samples to define all of his tectonic provinces. Furthermore his sample numbers for the various settings were as follows: oceanic island arc (9); continental island arc (9); active continental margin (2), and passive margin (3). As noted earlier all four of Bhatia's (1985) tectonic provinces are represented in the Welsh succession - an extremely unlikely occurrence given the history of the source area. It would, therefore, appear that there are a number of problems with his model, namely, (a) the relatively small sample size, particularly of some of the tectonic settings (e.g. passive margin); (b) the lack of recognition of the complexity of the geochemical history of the sediments used in creating his model; and (c) the use of certain elements, primarily Rb, Sr and Ba, which as noted above, show very poor correlation in terms of provenance as their distribution is affected by many variables.

## 5.10 Conclusions

The geotectonic setting for the southern part of the Welsh Basin has been discussed by a variety of authors (Bevins 1982; Bevins et al. 1984; Kokelaar et al. 1984a; Siveter et al. 1989; Woodcock in press). The current consensus, based largely on geochemical evidence, is that the Welsh Basin formed in an ensialic marginal basin in a continental margin which developed on the southern side of the Iapetus Ocean, subduction being initiated during the Tremadoc (Kokelaar 1979). The two dominant signatures, namely that of the pre-Tremadoc passive margin and the post-Tremadoc active continental margin, are supported by both the petrographic and geochemical analysis of the sediments deposited in the basin over a period of 30-35 Ma. The sediments, particularly the sandstones, reflect the longevity of the relict passive margin signature. This signature may be enhanced by the palaeogeographic position of the source area (southern Eastern Avalonia/Baltica) for the sandstone detritus. The mudstones are apparently more sensitive and thus reflect the active continental margin signature. There are serious flaws in certain geochemical provenance models for both sandstones and mudstones, as outlined above, and extreme caution should be exercised when applying them. Certainly they should not be used as the sole indicators of sedimentary provenance in areas which are poorly understood geologically.

The bimodal characteristics of the volcanic rocks in the source area are not truly reflected in many of the techniques. While this may be accounted for in framework grains by the easier weathering/alteration of basic fragments and their lower transport stability, it is of note that only a few of the geochemical techniques suggest any form of basic input. The most useful tools are the spider plot of trace elements and the K/Rb plot. In summary, given that the geology of provenance areas is commonly complex, it is best to use a variety of techniques, both petrographic and geochemical, in conjunction with other geological parameters to provide as complete a reconstruction as possible.

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## CHAPTER SIX - LLANDEILO-LLANDOVERY LAMINATED HEMIPELAGITES FROM THE WELSH BASIN

### Abstract

Laminated hemipelagites are recognised from Llandeilo-Llandovery age sediments within the Welsh Basin. X-ray diffraction and whole rock geochemical analyses failed to distinguish the hemipelagic sediments from associated mudstones and siltstone turbidites. Internally the hemipelagites are characterised by laminae of pyrite framboids. These laminae do not show any evidence of primary deposition and are interpreted as diagenetic products formed under euxinic or semi-euxinic conditions.

### 6.1 Introduction

Hemipelagic sedimentation is the background sedimentation within the majority of depositional basins. It occurs primarily by the slow settling of fine-grained sediment particles through the water column, although other processes, for example nepheloid layers and resuspension, may also have an effect (Eittreim 1984; Gorsline *et al.* 1984; McCave 1984; Stow & Piper 1984a). Turbidity current activity is commonly negligible. The organic-rich laminae are thought to originate by the vertical fallout of seasonal phytoplankton blooms, producing varve-like sediments, as noted from the California Borderland Basins (Isaacs 1984; Thornton 1984).

Hemipelagites are commonly homogeneous and structureless (Stow & Piper 1984b) although a depositional lamination may be preserved under anoxic conditions (Isaacs 1984; Thornton 1984). However, apart from their mixed biogenic-terrigenous aspect, it is difficult to generalise about their composition given the variations both in type and amounts of the biogenic and terrigenous inputs (Stow & Piper 1984b).

Laminated hemipelagites within turbidite facies have been documented by a number of authors from the Lower Palaeozoic Welsh Basin (e.g. Cave & Hains 1986; Dimberline 1987). Similar sediments have also been reported over a wide geographic area and stratigraphic range (c.f. Arthur et al. 1984; Robertson 1984; Stow & Piper 1984c). The laminated hemipelagites comprise silt-rich laminae alternating with organic carbon rich laminae. Based on X-ray diffraction analysis of the laminated hemipelagites and associated fine-grained turbidites, Dimberline (1987) subdivided the laminated facies into three types (see below) and proposed an elegant model to account for their distribution. This purpose of this paper is to test Dimberline's model for similar hemipelagic facies from the Llandeilo-Llandovery succession in west Wales and to show that the model does not appear to have wide applicability.

## 6.2 Location and stratigraphy

The study area is situated on the west coast of Wales and forms part of the Lower Palaeozoic Welsh Basin (Bassett 1984; Woodcock 1984; Siveter et al. 1989). In this area a succession of fourteen formations crop out, and range in age from Llandeilo (Nemagraptus gracilis Biozone) to upper Llandovery (Monograptus turriculatus Biozone) and young overall to the north (McCann & Pickering 1989).

The formations may be divided into two types based on the dominant lithology: (i) Sandstone-dominated formations (e.g. Newport Sands Formation, Aberystwyth Grits Formation) deposited mainly from high-concentration turbidity current flows. They may be classified according to the facies scheme of Pickering et al. (1986) as gravelly muds (Facies Class A), sandstones (Facies Class B) and sand-mud couplets (Facies Class C). (ii) Mudstone-dominated formations (e.g. Gwbert Formation, Tresaith Formation) comprise thin silts, muddy silts and silt-mud couplets (Facies Class D) which are laterally variable. The mudstone and siltstone turbidite beds are commonly overlain by bioturbated pelagic muds

(Facies Class E) with well-laminated hemipelagites being locally abundant (e.g. Tresaith Formation, Gaerglwyd Formation). Bioturbation is absent in the laminated hemipelagite facies.

The distribution of the dominant lithologies appears to be determined by global eustatic sea-level changes in the area. For the strata of the study area, two major periods of transgression have been recognised, one in the Caradoc and the second in the early Llandovery (McKerrow 1979; Leggett *et al.* 1981). These are separated by a eustatic regression in the late Ashgill which is related to a major glacial event in Gondwana (Brenchley & Newell 1984). Fortey & Cocks (1986) noted that the lithological and faunal changes in parts of Wales to the south of the study area may be related to the global eustatic curve and suggested that it was not necessary to invoke a tectonic influence. All of the sandstone-dominated formations (with minor exceptions - see Chapter Four) may be related to these eustatic sea-level changes, with maximum turbidity current activity occurring during periods of transgression and regression and at periods of low sea-level stand.

### 6.3. Laminated hemipelagites

A variety of origins have been proposed for laminated hemipelagic sediments. These range from the laminae being annual varves due to phytoplankton blooms (Thornton 1984; Warren *et al.* 1984) or part of the Bouma (1962) Td division (Carey & Roy 1985). Aggregate structures within the silt laminae have been variously interpreted as faecal pellets of planktonic or benthonic organisms, or compactional features (Dimberline 1987). In thin section the hemipelagic lithofacies, of the study area, is composed of alternating light-coloured, quartz-rich and dark-coloured, organic- and clay-rich laminae.

Results from a number of Deep Sea Drilling Project surveys have revealed that authigenic dolomite is commonly found in

continental margin and deep-sea hemipelagic sediments of Miocene to Holocene age (e.g. Matsumoto 1983; Shimmield & Price 1984; Lumsden 1988). This knowledge, coupled with X-ray diffraction analysis of 66 samples, led Dimberline (1987) to suggest that the Wenlock age laminated hemipelagites and associated turbidites of the Welsh Basin could be divided into three distinct types and one sub-type. These divisions were recognised on the basis of the presence or absence of certain carbonate and sulphide minerals. Type 1 contained no calcite, dolomite or pyrite. Type 1a contained calcite only. Type 2 contained calcite, dolomite and pyrite, while Type 3 contained only dolomite and pyrite. Analysis of the dolomite rhombs suggested that at least a proportion of them may have formed in situ by displacement and/or replacement of the host sediment at shallow depth in anoxic, organic-rich marine sediments (Dimberline 1987).

Dimberline (1987) formulated a model to account for the three different hemipelagic and turbiditic mudstones recognised. Given that each of the recognised types contained a diagnostic assemblage of minerals then they could be related back to the sedimentary inputs into the system. These included the presence or absence of reactive organic matter, reactive iron hydroxides and detrital calcite. The conditions, for example, low pH and sulphate reduction, under which these various inputs would then be diagenetically altered to produce the characteristic assemblages, could then be determined.

#### **6.4 X-ray diffraction analysis of Llandeilo-Llandovery sediments**

Thirty-six samples were chosen for x-ray diffraction analysis from the coastal succession of west Wales. These were analysed by means of a Phillips X-ray diffraction analyser. The results of this analysis are presented in Fig 6.1. None of the samples contained any calcite or dolomite above detectable limits (approximately 3.0%) although some did contain pyrite. All of the samples contain quartz, chlorite,

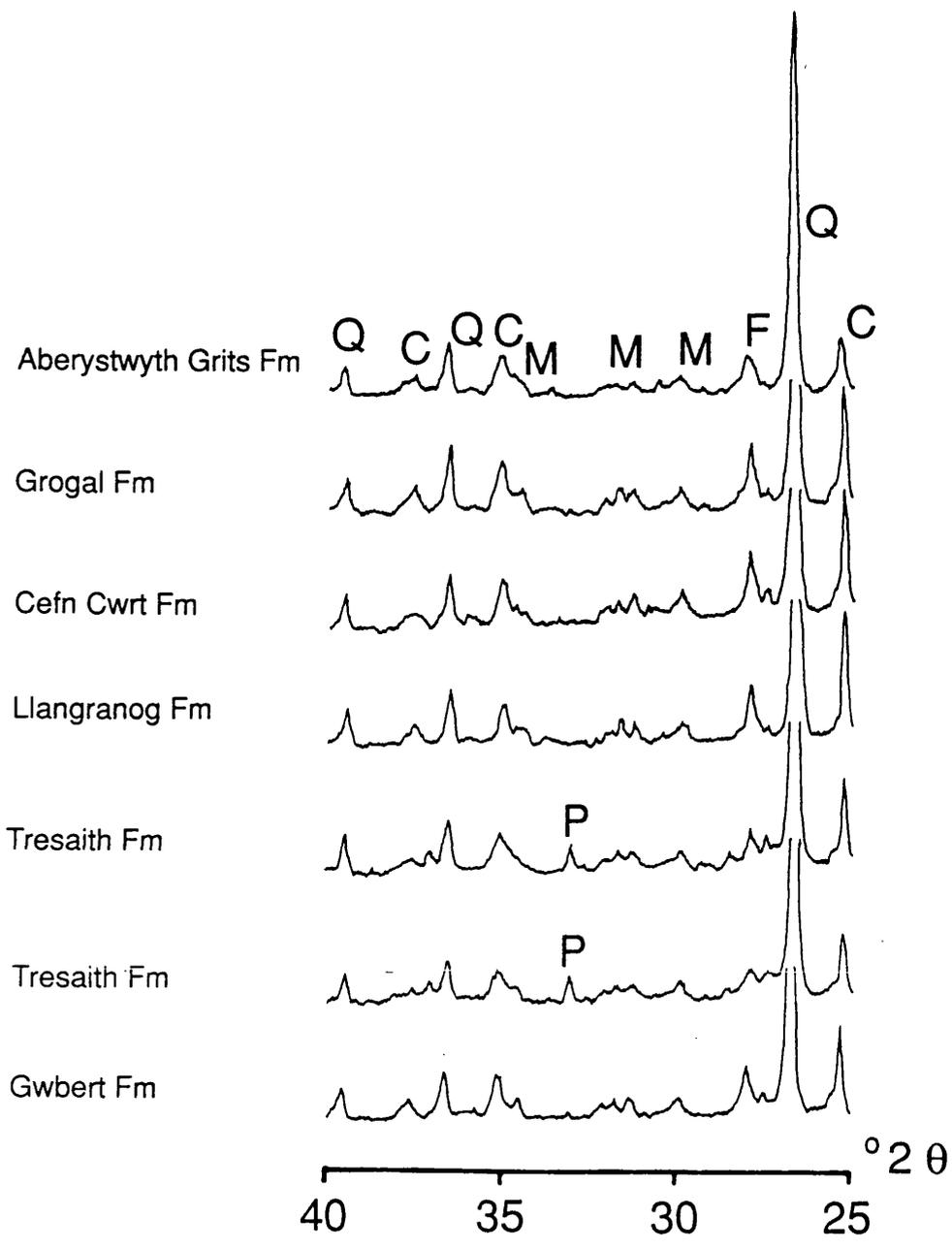


Figure 6.1 Selected typical bulk-rock XRD traces for the sediments studied. Key: Q - Quartz, F - Feldspar, C - Chlorite, M - Mica, P - Pyrite.

and mica and are very similar to Dimberline's samples except for the absence of carbonates.

Following on from the initial XRD runs, a further twenty samples were chosen for more detailed analysis. This involved both longer XRD runs (5-50 2 $\theta$ ) and also smaller sample intervals. The initial set of thirty six samples were spread over a wide area both stratigraphically and geographically and it was, therefore, considered appropriate to constrain the sample interval somewhat. A series of samples were taken from tripartite sedimentary sequences comprising a basal silty layer overlain by a dark turbiditic mudstone followed by a lighter coloured oxic, bioturbated mudstone. These tripartite sequences are common in the study area and are the dominant facies types in some formations (e.g. Gwbert Formation, Tresaith Formation). The XRD traces for these sequences reveal that there are no real differences between the three layers apart from the intensity of the quartz peak (Figs. 6.2, 6.3). The chlorite-mica peaks are also variable although to a lesser extent.

A further series of samples was taken from a sequence comprising thin laminated hemipelagites and mudstone turbidites (Fig. 6.4). Similar sequences may be found throughout the succession. Again, there are no real differences between the various layers, quartz peak intensity being the only significant variable.

### 6.5 Levels of organic carbon

The levels of total organic carbon were measured on a Carbon-Sulphur Determinator in thirty five samples from the succession. The results of these analyses, however, were not definitive. The light-grey bioturbated mudstones showed the lowest levels of total organic carbon (0.0-0.0058%). Percentage organic carbon did appear to be related to the degree of bioturbation since one sample from sediments with very rare bioturbation gave a reading of 0.201%. The dark-grey mudstones showed the highest recorded levels of TOC

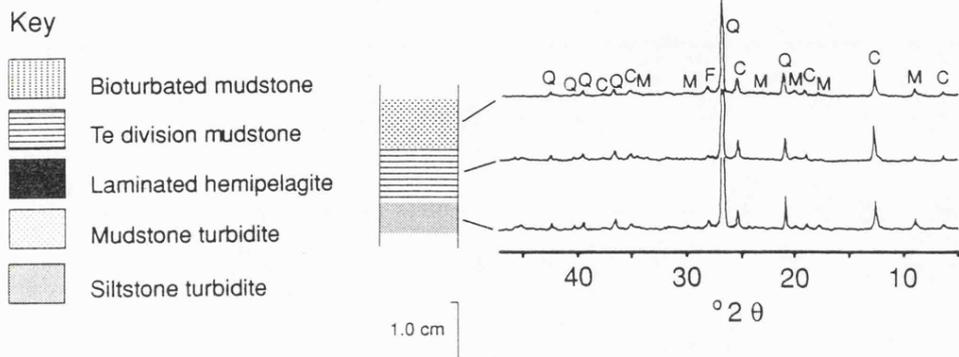


Figure 6.2 Bulk-rock XRD traces for silt-mud tripartite sequences from the Gwbert Formation, Gwbert. Key as for Fig. 6.1.

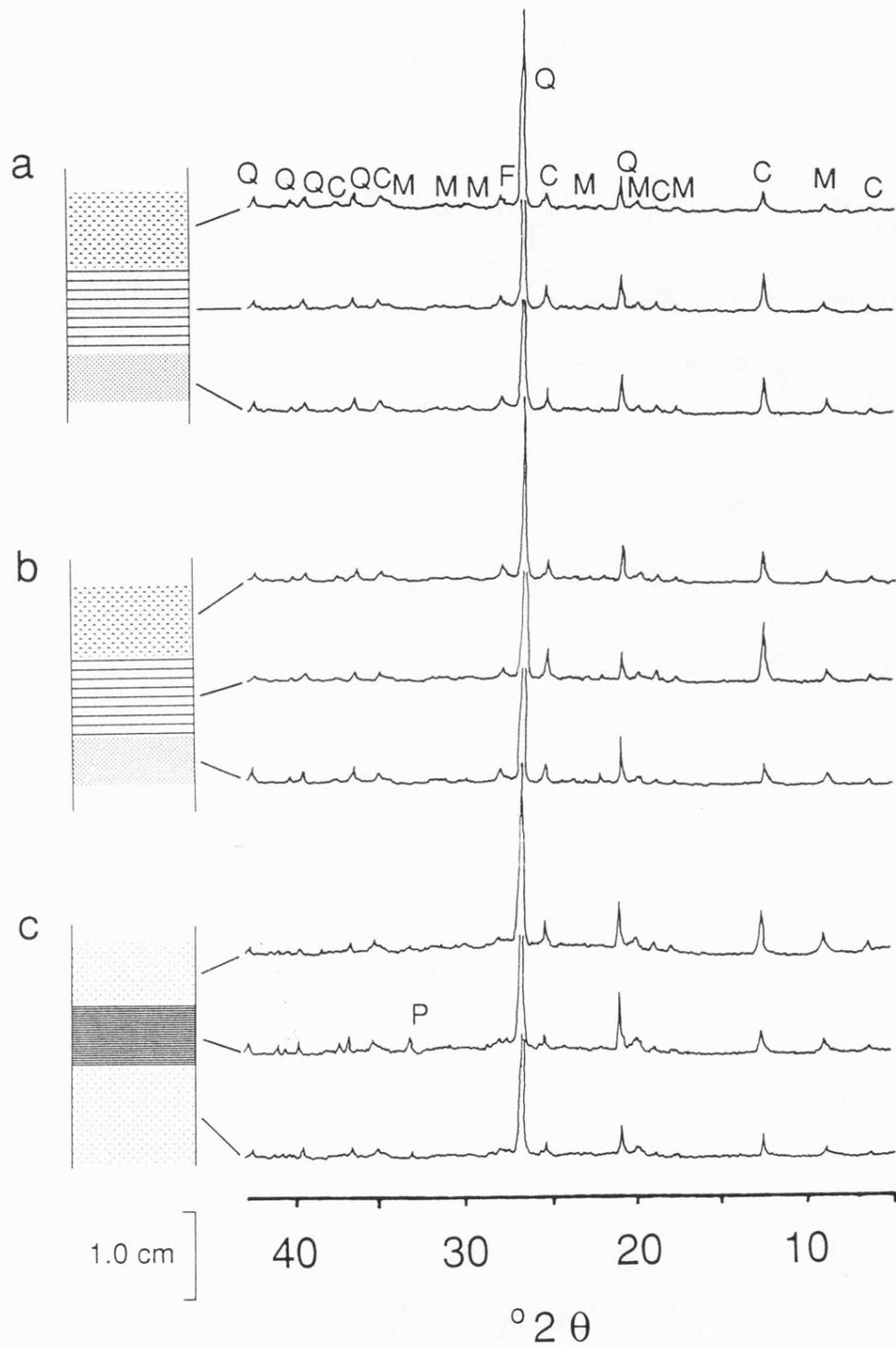


Figure 6.3 Bulk-rock XRD traces for silt-mud tripartite sequences from the Tresaith Formation, Tresaith. Key as for Fig. 6.1.

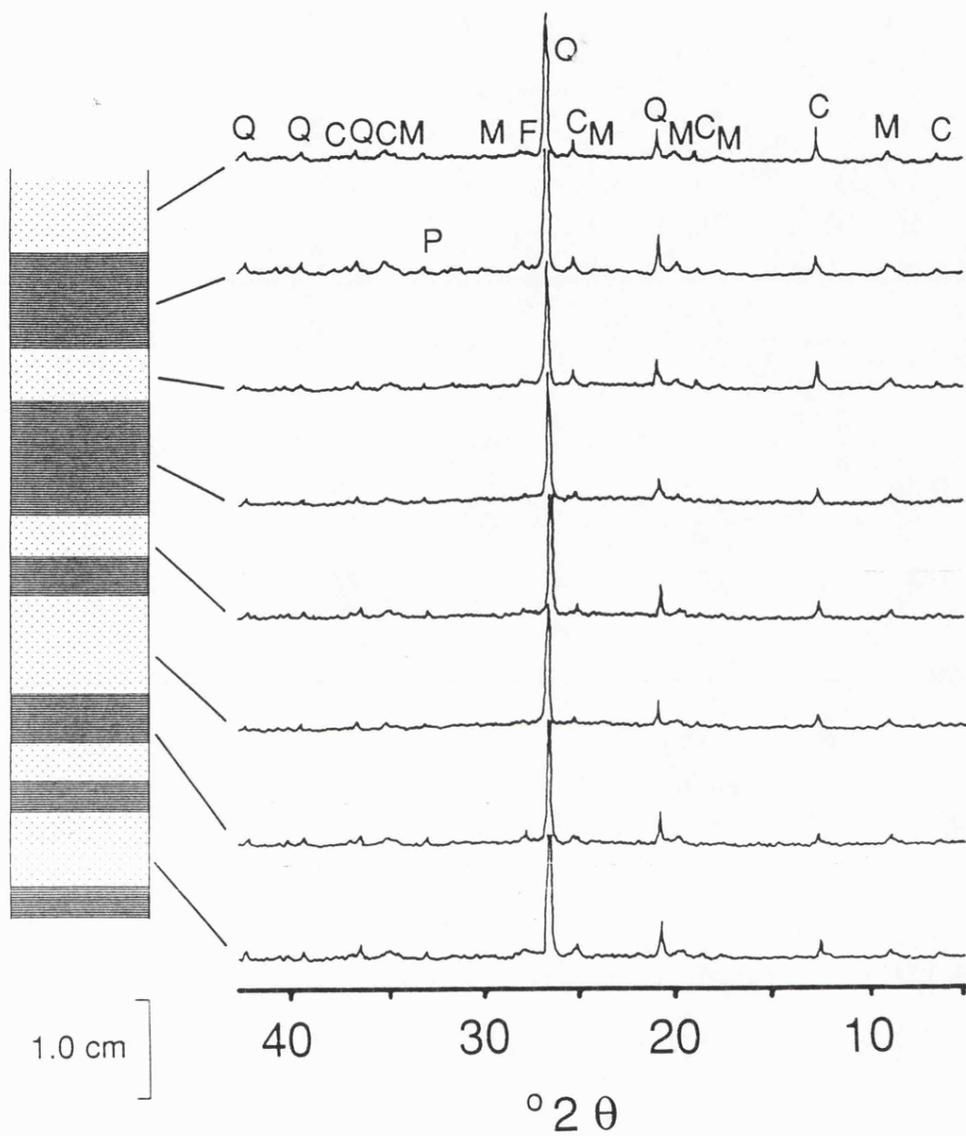


Figure 6.4 Bulk-rock XRD traces for mudstone turbidites and laminated hemipelagites from the Gaerglwyd Formation, Traeth-yr-Ynys. Key as for Fig. 6.1.

(0.037-0.276%) while the laminated hemipelagites showed intermediate values (0.004-0.117%). However, all of these values are low, suggesting that the biogenic layers were not very organic rich.

#### 6.6 Geochemistry of hemipelagites and fine-grained turbidites

Dimberline (1987) presented geochemical analyses for his three deep-sea deposit divisions and noted that Type 1 had significantly less CaO and slightly more  $\text{Al}_2\text{O}_3$ ,  $\text{Fe}_2\text{O}_3$  and  $\text{K}_2\text{O}$  than Types 2 and 3 (Table 6.1). Whole rock major and trace element analysis of 29 samples from the coastal succession reveals that there is very little variation within those elements which are deemed to be important by Dimberline (1987). The levels of CaO are low in all of the samples except for a single sample from the Grogal Formation which is 1.1% (Table 6.1). This is still much lower than the figures of 3.36% and 4.11% given by Dimberline (1987) for Types 2 and 3 respectively. Similarly the values of  $\text{Al}_2\text{O}_3$ ,  $\text{Fe}_2\text{O}_3$  and  $\text{K}_2\text{O}$  are all higher than the ranges suggested by Dimberline (1987).

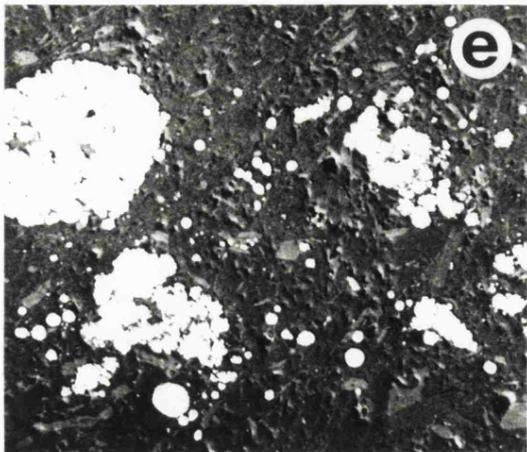
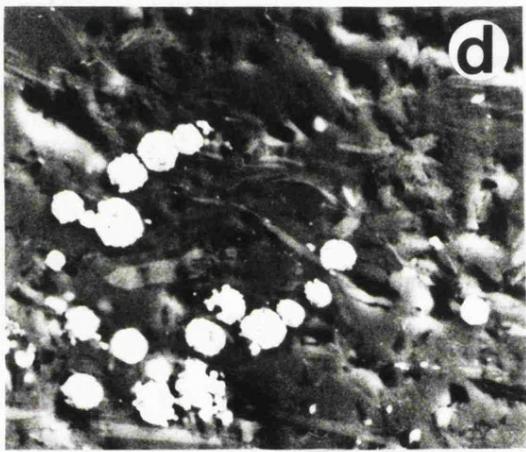
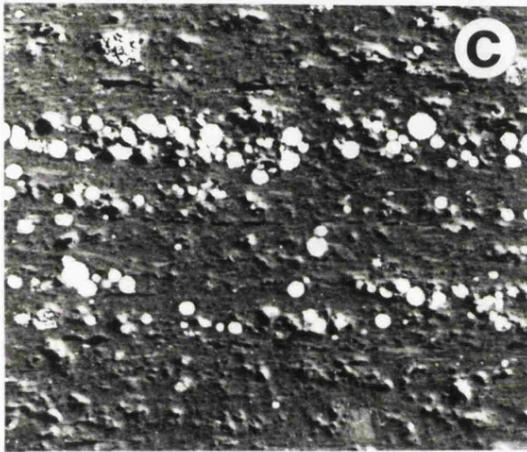
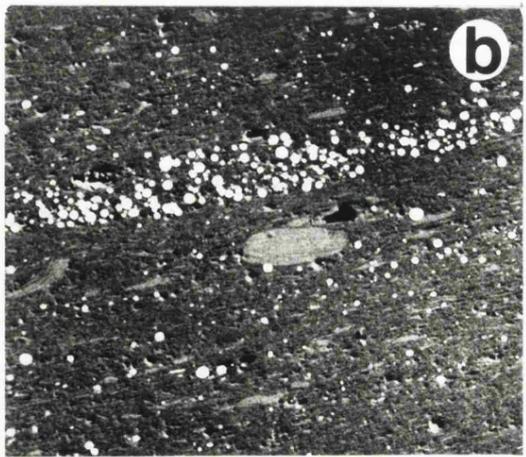
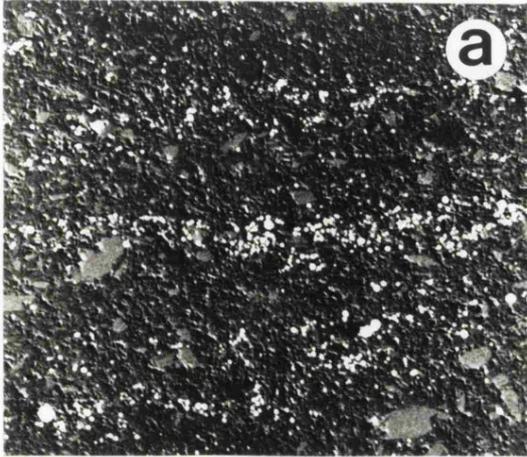
#### 6.7 Back-scattered electron microscopy of hemipelagites and associated fine-grained turbidites

The scanning electron microscope was used in backscatter mode to examine 23 samples from the succession (Plate 6.1). The samples were distributed over a wide stratigraphical area and represent all the dominant mudstone lithologies present in the study area. Only framboidal pyrite was recognised.

Framboidal pyrite is most common within the laminated hemipelagites. The framboids are circular (spherical when examined using the SEM - Pl. 6.1 f) in section and range in size from 2  $\mu\text{m}$  to 150  $\mu\text{m}$ , the larger ones tending to be isolate. They are either scattered throughout the sediment or else they form thin laminar structures (Pl. 6.1 a, b, c), arranged parallel to bedding, composed of adjoining pyrite

## Plate 6.1

- a. BSEM image of laminar arrangement of pyrite framboids in anoxic hemipelagites of the Gaerglwyd Formation (Lower Llandovery), Traeth-yr-Ynys, x 120.
- b. BSEM image of laminar pyrite framboids from the Parrog Formation (Llandeilo), Parrog, x 120.
- c. BSEM image showing details of individual laminae from the Parrog Formation (Llandeilo), Parrog. Note their fact that they are often only one framboid thick, x 240.
- d. BSEM image showing detail of circular arrangement of pyrite framboids from the Gaerglwyd Formation (Lower Llandovery), Traeth-yr-Ynys, x 900.
- e. BSEM image showing details of individual pyrite framboids, Tresaith Formation (Ashgill), Tresaith, x 180.
- f. SEM image showing details of individual pyrite framboids and their spherical nature in clays of the Tresaith Formation (Ashgill), Tresaith, x 420 (upper portion).



framboids. These laminae may be up to 130  $\mu\text{m}$  thick and are laterally continuous over a few centimetres. It should, however, be noted that framboidal pyrite may also be rare and isolate in some parts of the laminated hemipelagite facies.

Mudstone and siltstone turbidites associated with the laminated hemipelagites also contain framboidal pyrite with the former containing the larger concentration. Framboids within these sediments are most commonly isolated although they may exhibit some planar arrangement. Clumps of framboids may be locally present. Within the mudstones the framboids may be concentrated towards the base of the graded unit.

Similarly mudstone and siltstone turbidites from areas where bioturbation levels are high have very low levels of pyrite. In these sediments the framboids tend to be isolate although rare occurrences of framboid laminae arranged parallel to bedding have been noted.

## 6.8 Discussion

Dolomite is a widespread, albeit minor, component of deep-marine and continental margin sediments (Baker & Burns 1985). It is found in the majority of ocean basins, particularly smaller ones, throughout post-Jurassic time, and the amounts of dolomite (averaging 1% - global DSDP samples) are comparable to those in modern supratidal and sabhka carbonate settings (Lumsden 1988). Most deep-marine dolomite formation was initiated at shallow burial depths (0.0-1.0 m) shortly after deposition. The necessary magnesium was derived from pore waters that diffused and advected in from the overlying seawater. The calcium, carbon and oxygen were derived locally by organic maturation and carbonate recrystallisation (e.g. Baker & Burns 1985).

Two of Dimberline's (1987) sediment divisions contained dolomite. Type One, however, does not and he suggested a number of reasons, including high sedimentation rates and low

rates of calcium carbonate dissolution, to account for this situation. Two reasons, however, were considered to be of greater importance: (i) a low reactive organic matter content, possibly as a result of sulphate reduction in the water column; the sulphate reduction of organic matter produces the high  $\text{HCO}_3^-/\text{Ca}^{2+}$  ratio needed for dolomite precipitation, and, (ii) a low reactive iron content as a result of the weathering in the source rocks; the input of reduced iron would have raised pH levels and favoured carbonate precipitation.

It should be noted that the above reasons are to account for areas in which no calcite, dolomite or pyrite occur. For types 2 and 3, where dolomite is present, it is always found in conjunction with pyrite. In the Llandeilo and Llandovery sediments of west Wales, however, pyrite is locally abundant but there is no evidence, from XRD analysis, for the presence of dolomite. This would suggest that in the study area the controls on the system are quite different from those suggested by Dimberline (1987) for the Wenlock sediments of the Welsh Basin.

The criteria used by Dimberline (1987) to distinguish his various subdivisions were based on XRD analysis, back-scattered electron microscopy and whole rock geochemistry, with the former being considered the most important. He did not, however, give details as to how many samples from the original total were used to define each individual division.

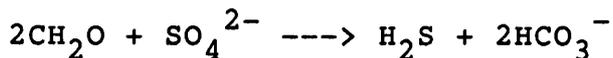
X-ray diffraction analysis of the Llandeilo-Llandovery samples reveals that dolomite and calcite are completely absent, or at least are below the detection limits of the method. Indeed the only mineral from Dimberline's (1987) classificatory scheme represented in the samples is pyrite. As noted above, in the Wenlock samples dolomite and pyrite XRD peaks can be correlated suggesting that their origins, or at least the physical and chemical conditions necessary for their formation, were similar. The absence of dolomite in the sediments of the study area would suggest that the

controls on the system must be different.

The more detailed XRD traces reveal that the intensity of the quartz peaks, and to a lesser extent the chlorite peaks, were the only features distinguishing laminated hemipelagites, turbiditic mudstones and mudstone and siltstone turbidites. These traces also confirmed the absence of carbonate minerals and the presence of pyrite within the laminated hemipelagites.

Dimberline (1987) suggested two main reasons as to why dolomite might not occur within sediments (see above). These reasons were partly determined by his observed pyrite and dolomite co-existence and thus any explanation would need to account for the absence of both minerals. The reasons given were also determined by the then current state of knowledge on deep-sea dolomites. More recently, however, Lumsden (1988) has suggested that there is no relationship between the presence of organic matter and dolomite formation. In the sediments of this study, the presence of pyrite and the absence of dolomite suggests that no such explanatory constraints need to be invoked. Instead, physical and chemical factors which might preclude the formation of dolomite and yet permit the formation of pyrite must be considered.

Pyrite forms during shallow burial via the reaction of detrital iron minerals with  $H_2S$  (Goldhaber & Kaplan 1974; Berner 1984).  $H_2S$  is produced by the reduction of interstitial dissolved sulphate by bacteria using sedimentary organic matter as a reducing agent and energy source. This may be summarised by the following equation:

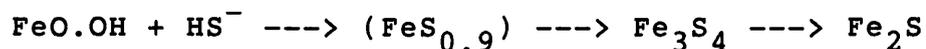


The reaction proceeds via a number of intermediate stages and can only operate under anoxic conditions (Davis *et al.* 1988). The main controls on the reaction are: (a) the amount of reactive organic matter, and, (b) the availability of dissolved sulphate. Pyrite formation is also limited by the

amount and reactivity of detrital iron minerals. Indeed the amount formed is more dependent on this than on the amount of organic matter.

As noted previously, all of the pyrite documented in the sediments of the study area is framboidal in form. It was originally thought that framboids of pyrite necessitated spheroidal precursors but experiments have proved these to be unnecessary (Raiswell 1982). Instead, the origin of the framboids is dependent on chemical factors which influence the extent to which certain metastable iron monosulphide intermediates (i.e. greigite and mackinawite) are involved.

There are a variety of pathways by which framboid development may proceed. Initial iron sulphide precipitates (sometimes identified as mackinawite) develop a spheroidal texture on transformation to greigite. This texture could then be maintained or else changed to a framboidal one by internal nucleation of pyrite crystals. This may be summarised by the following reaction pathway:



Pyrite form, therefore, is controlled by the type of iron monosulphide intermediates generated which, in turn, are determined by the sediment and pore water compositions (Raiswell 1982).

As noted earlier the framboids may be arranged in laminae within the hemipelagic sediments. Similar pyrite laminae have been documented from the Cambrian in Sweden (Thickpenny 1984) where they were interpreted to be a primary sedimentary deposit. In the sediments of the study area, however, there is no evidence of any primary sedimentary features or associated wet-sediment deformation features within the pyrite laminae (sensu Schieber 1989). This would suggest that the mechanism producing the laminae is a diagenetic rather than a depositional one.

Morris (1980) has suggested that where sedimentation rates

are low and the oxic/anoxic boundary is close to the surface then pyrite framboids form mainly from the iron available within the sediment. Microscopic scale redistribution of in situ iron would lead to the formation of framboids while transport of iron over a distance of centimetres would result in the formation of euhedral pyrite. A number of authors have suggested methods, commonly involving faunal and sedimentological criteria, by which the presence of euxinic or semi-euxinic depositional conditions can be deduced from ancient environments (e.g. Morris 1979; Arthur et al. 1984; Savrda & Bottjer 1987; Raiswell & Berner 1985). The continuous lamination, absence of bioturbation and benthic fauna and the sediment colour would all suggest that the laminated hemipelagic sediments of the study area were all deposited under conditions of little or no oxygen.

In euxinic environments,  $H_2S$  is present both in the water column and within the sediment (Raiswell & Berner 1985). This facilitates the formation of syngenetic pyrite. Of greater importance, however, is the fact that organic carbon is not required at a given location for pyrite formation, instead the amount of reactive detrital iron is of greater importance (Raiswell 1982; Berner 1984; Raiswell & Berner 1985). This is very much the situation in Recent sediments (Westrich & Berner 1984).

Given that the hemipelagic sediments of the study area were deposited under euxinic or semi-euxinic conditions (based on the criteria outlined above), then the pyrite present could have formed both syngenetically and diagenetically. The low values of organic carbon recorded within the laminated hemipelagites, however, suggests that the production of sulphide from seawater sulphate by the bacterial oxidation of organic matter was an important reaction and that much of the pyrite was produced diagenetically. Some of the laminae may have been composed of syngenetic pyrite but the greater part of the framboids were probably formed as a result of a redistribution of iron on a microscopic scale within the sediment (e.g. Morris 1980).

Excess iron was probably supplied by terrestrial runoff in colloidal form from the major river deltas which lay to the south and east of the depositional area throughout much of the Llandeilo - Llandovery. Flocculation of the iron hydroxides would have delivered relatively large and concentrated units of reactive iron into the basin. These flocs were then either incorporated into turbiditic sediments where they would have a random distribution or else within hemipelagic sediments where they would be parallel laminated. This would account for the clumps of pyrite framboids noted in many of the turbiditic sediments. The presence of concentrations of pyrite framboids at the base of some of the mudstone turbidites does suggest a depositional origin for at least a proportion of the framboids. These may have formed syngenetically.

Finally, it should be noted that there are at least two other reasons which may be used to explain the absence of carbonate from the sediments in the present study. Firstly, calcareous nannoplankton did not evolve until the early Jurassic (and were thus not a source for detrital carbonate) and secondly it is entirely possible that the sediments were deposited below the Carbonate Compensation Depth (CCD), thus precluding the formation of carbonate. Either of these reasons would also cast doubts on the widespread applicability of Dimberline's classification.

## 6.9 Conclusions

(1). The Dimberline (1987) model of predicting depositional setting based on the presence or absence of certain carbonate minerals and pyrite in x-ray diffraction traces clearly has only limited usage. There was no evidence of any such divisions within the samples from the present study area. This is despite the fact that the number of samples for which XRD traces were determined was almost equal to the total number of samples used by Dimberline (1987) to construct his model in the first instance.

(2). XRD traces taken from closely-spaced samples revealed that there was very little difference between the various facies types. Laminated hemipelagites and siltstone and mudstone turbidites varied only in the height of the quartz peak.

(3). Whole rock geochemical analysis of 29 samples of laminated hemipelagites, turbiditic Te (Bouma 1962) divisions and mudstone turbidites showed that they were virtually indistinguishable. None of the criteria outlined by Dimberline (1987) for differentiating them could be applied.

(3). Pyrite, where present, was always framboidal in form. This would suggest that the reactive iron was only transported on a microscopic scale within the sediment.

(4). Laminae of framboidal pyrite observed in the hemipelagites do not show any primary depositional features. Nor do they exhibit any characteristics consistent with wet-sediment deformation. These two factors would suggest that the laminae are a diagenetic rather than a depositional feature.

(5). Sedimentological and faunal criteria suggest that the sediments were produced under euxinic or semi-euxinic conditions. Under these conditions, the main control on pyrite formation is the presence or absence of reactive iron. Iron hydroxides were derived both from the continental landmass via the rivers and also from the erosion and weathering of coeval and older volcanic rocks all of which lay to the south and east of the depositional area.

(6). Given that diagenetic dolomites are found in many deep-sea and continental margin areas, the controls operating on the sediments deposited from Llandeilo-Llandovery times must have been such as to preclude dolomite formation while still allowing pyrite to form. Such factors would have included: (i) high sedimentation rates with low rates of inward  $Mg^{2+}$  diffusion; (ii) low rates of detrital calcium carbonate dissolution; (iii) the  $Ca^{2+}/HCO_3^-$  ratio not

favouring dolomite formation; and (iv) deposition below the Carbonate Compensation Depth. All of these factors would have prevented the precipitation of dolomite without influencing the formation of pyrite.

(7). It is also possible that the lack of carbonate minerals is a function of either the lack of calcareous nanoplankton to provide a detrital carbonate source or else that the sediments were deposited below the Carbonate Compensation Depth.

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CHAPTER SEVEN - DISTRIBUTION OF ORDOVICIAN-SILURIAN  
ICHOFOSSIL ASSEMBLAGES IN WALES - IMPLICATIONS FOR  
PHANEROZOIC ICHNOFAUNAS

**Abstract**

The Ordovician-Silurian stratigraphic succession of the coastal outcrop of west Wales consists of two dominant lithologies, namely sandstone and mudstone. The distribution of these lithologies is largely determined by eustatic sea-level changes. The strata contain a diverse and relatively abundant ichnofaunal assemblage consisting of fifteen ichnogenera: Chondrites, Circulichnis, Cochlichnus, Cosmorhapse, Desmograpton, Gordia, Helminthoida, Helminthopsis, Nereites, Neonereites, Palaeophycus, Paleodictyon (Glenodictyon), Paleodicyton (Squamodictyon), Planolites, Protopaleodictyon and Spirophycus. The ichnogenera are unevenly distributed throughout the succession, the main controlling factors on their distribution being toponomy, anoxia, and a global extinction event. The relative importance of these controls is examined with reference to 33 previously described and taxonomically well-documented deep-water flysch ichnofossil assemblages covering the Cambrian to Tertiary worldwide.

**7.1 Introduction**

There are many problems associated with the analysis of the controls on ichnofossil distribution. A number of factors, either in isolation or together, may affect the ichnofossil distribution and a variety of controls have been identified (e.g. Seilacher 1967, 1977) in the quest to aid palaeoenvironmental interpretations. This study aims to reassess this earlier work by critically examining how useful these existing distributional models are for aiding the

interpretation of a number of ichnofaunal assemblages through a well-exposed coastal succession of Ordovician-Silurian age in west Wales. The area is ideally suited for testing the various hypotheses, as it shows a continuous succession with a resolved stratigraphy that includes the Ordovician-Silurian boundary. Ichnofossil assemblages are reasonably common throughout the sequence in question.

## 7.2 Location, Lithologies and Age of Study Area

The study area is situated on the west coast of Wales and forms part of the Lower Palaeozoic Welsh depositional Basin, the possible tectonic setting of which has been the subject of a number of reviews (e.g. Bassett 1984; Woodcock 1984; Siveter et al. 1989). The area comprises a succession of fourteen formations which range in age from the Ordovician Llandeilo Series (Nemagraptus gracilis Biozone) to the Silurian upper Llandovery age (Monograptus turriculatus Biozone) and consistently young to the north (McCann & Pickering 1989).

The formations may be subdivided into two types based on the dominant lithology. Sandstone-dominated formations (e.g. Newport Sands Formation, Aberystwyth Grits Formation) are deposited from high-concentration turbidity current flows. They may be classified according to the facies scheme of Pickering et al. (1986) as gravelly muds (Facies Class A), sandstones (Facies Class B) and sand-mud couplets (Facies Class C). Mudstone-dominated formations (e.g. Gwbert Formation, Tresaith Formation) comprise thin silts, muddy silts and silt-mud couplets (Facies Class D) which vary considerably in lateral continuity over outcrop distances up to tens of metres. The mudstone-dominated formations show a wide range of features consistent with deposition from dilute turbidity currents. The turbiditic beds are frequently overlain by bioturbated pelagic muds (Facies Class E) with laminated hemipelagites being locally abundant (e.g. Gaerglwyd Formation).

The distribution of the dominant lithologies appears to be determined by eustatic sea-level changes in the area (Fig. 7.1). For the rocks in question two major periods of transgression have been recognised, in the Caradoc and the early Llandovery respectively (McKerrow 1979; Leggett *et al.* 1981). These are separated by a eustatic regression in the late Ashgill which is related to a major glacial event in Gondwana (Berry & Boucot 1973; Brenchley & Newall 1984). All of the sandstone-dominated formations may be related to these eustatic sea-level changes with maximum turbidity current activity occurring during periods of low sea level stand.

### 7.3 Ichnofossil Distribution in the Ordovician–Silurian succession of west Wales

The current research has recognised fifteen ichnogenera from the study area. A detailed systematic examination of the ichnofossils is presented in Chapter Eight. The recorded ichnogenera include Chondrites ichnospp., Circulichnis ichnospp., Cochlichnus ichnospp., Cosmorhapse ichnospp., Desmograpton ichnospp., Gordia ichnospp., Helminthoida ichnospp., Helminthopsis ichnospp., Neonereites ichnospp., Nereites ichnospp., Palaeophycus ichnospp., Paleodictyon (Glenodictyon) ichnospp., Paleodictyon (Squamodictyon) ichnospp., Planolites ichnospp., Protopaleodictyon ichnospp. and Spirophycus ichnospp. Both Desmograpton and Spirophycus are recorded for the first time from the Palaeozoic (McCann 1989). A number of observations may be noted about the stratigraphic distribution of the ichnofossils within the study area (Fig. 7.2):

a) Ichnofossils are present in all of the formations except for the Ceibwr Formation. The probable reason for this is that the mudstone-dominated Ceibwr Formation has been extensively sheared, giving rise to a penetrative cleavage that has destroyed any trace of bioturbation.

b) Chondrites ichnospp. and Planolites ichnospp. are the most

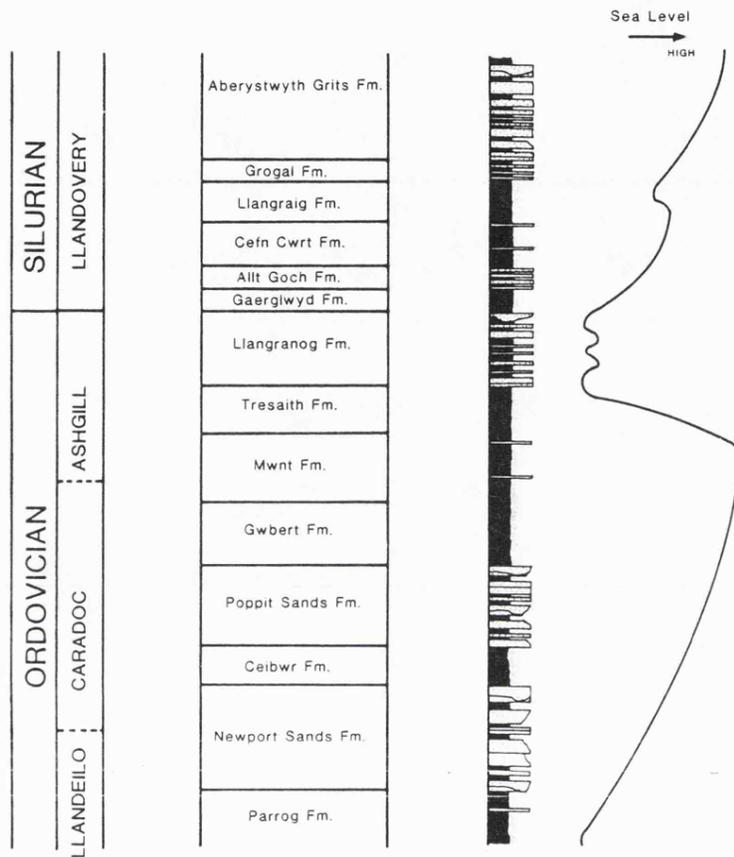


Figure 7.1 Generalised stratigraphic column of the area relating the lithological divisions to eustatic sea-level changes. Sea-level curve after McKerrow (1979) and Leggett *et al.* (1981).

AGE	ORDOVICIAN							SILURIAN						
FORMATION	Parrog	Newport Sands	Ceibwr	Poppit Sands	Gwbert	Mwnt	Tresaith	Llangranog	Gaerlwyd	Allt Goch	Cefn Cwrt	Llangraig	Grogal	Aberystwyth Grits
<i>Chondrites</i>		●		●	●	●	●	●	●	●	●	●	●	●
<i>Circulichnis</i>								◐						◐
<i>Cochlichnus</i>								○						◐
<i>Cosmorhapse</i>														○
<i>Desmograpton</i>														○
<i>Gordia</i>								◐						●
<i>Helminthoida</i>														○
<i>Helminthopsis</i>				●				●	●					●
<i>Nereites</i>														○
<i>Neonereites</i>														○
<i>Palaeophycus</i>		●		●				●	●					●
<i>Paleodictyon (Gleno.)</i>														●
<i>Paleodictyon (Squamo.)</i>														●
<i>Planolites</i>	○	●		●	●	●	●	●		●	●	●	●	●
<i>Protopaleodictyon</i>								◐						●
<i>Spirophycus</i>														○

Figure 7.2 Stratigraphic distribution and abundance data of the sixteen ichnogenera recorded from the Welsh coastal succession. The symbols refer to numbers of observed specimens.

commonly recorded ichnofossils in the succession. In mudstone-dominated formations they are the only recorded ichnofossils.

c) Ichnofaunal assemblages change both in abundance and diversity through the succession. This is particularly obvious when examining the changes from the older (i.e. Newport Sands Formation) through to the younger (i.e. Aberystwyth Grits Formation) sandstone-dominated formations.

d) There is a decrease in total numbers of ichnogenera at the Ordovician-Silurian boundary.

### 7.3.1 Recurrent ichnofossil associations

Out of the fifteen recorded ichnofossils, 2 main associations (sensu Brenchley and Cocks 1982) have been recognised: the Chondrites-Planolites association, and the Graphoglyptid association.

The Chondrites-Planolites association comprises Chondrites and/or Planolites and is found in all of the formations except for the Ceibwr Formation (see above). The association predominates in the mudstone-rich formations. Both Chondrites and Planolites are considered to be facies-crossing ichnogenera with a wide environmental tolerance.

The Graphoglyptid association comprises burrows and trails of surface and/or sub-surface-dwelling organisms which were eroded and cast by passing turbidity currents. As such the distribution of the association is closely linked to the presence or absence of turbidite beds. The association comprises typical graphoglyptids and related ichnofossils, including, Circulichnis, Cochlichnus, Cosmorhaphe, Desmograpton, Gordia, Helminthoida, Helminthopsis, Nereites, Neonereites, Paleodictyon (Glenodictyon), Paleodictyon (Squamodictyon), Protopaleodictyon and Spirophycus (see Seilacher 1977 for further details).

### 7.3.2 Palaeoenvironments and trace fossil associations

The Chondrites-Planolites association is an environmentally tolerant one which is present in almost all of the formations in the study area. It is particularly dominant in the mud-rich formations which were deposited from dilute turbidity currents as fan-fringe, and lobe-fringe depositional environments. Associated hemipelagic sediments were deposited as background sedimentation in the basin. The Graphoglyptid association, however, was much more environmentally restricted (see below). It is confined to those sediments deposited on sandstone lobes and in channel-margin and interchannel environments.

### 7.4 Factors affecting the ichnofossil distribution

Controlling factors on ichnofossil distribution have been discussed by a number of authors (e.g. Seilacher 1967; Pickerill et al. 1984). Indeed, Ekdale (1988) has suggested that the distribution of trace-making organisms (and by extension trace fossils) in subaqueous environments is largely determined by three primary factors, i.e. water depth, water chemistry and substrate. A further factor, which could be significant in the west Wales succession is that of global extinction events, given that the Ordovician-Silurian boundary crops out in the area. These four factors, therefore, have been examined to explain the observed patterns of occurrence of ichnogenera in the study area.

#### 7.4.1 Toponymy

Toponomic preservation has been recognised as an important factor governing the distribution of ichnofossils within a succession (Pickerill et al. 1984), given that the fossil is a partial record of both the activity of the trace-making organism and of the type of substrate in which it lived (Frey & Seilacher 1980). In the present area the diversity and

abundance of ichnofossils is much lower in the mudstone-dominated formations (e.g. Gwbert Formation, Cefn Cwrt Formation) relative to the sandstone-dominated formations (e.g. Llangranog Formation, Aberystwyth Grits Formation) (Fig. 7.2). As noted above, Chondrites and Planolites are the only recorded ichnofossils in mudstone-dominated formations. The most obvious factor controlling such occurrences would appear to be the presence or absence of coarser-grained beds which would act as potential preservational sites for ichnofossils.

From a total of 16 ichnogenera preserved in the succession 12 (75.0%) are only ever preserved hypichnially while a further 2 (12.5%) are only ever preserved epichnially (sensu Martinsson 1970). The distribution of these ichnogenera is therefore very strongly controlled by the presence or absence of sandstone or siltstone beds which would preserve the trails. This leaves just two ichnogenera, namely Chondrites and Planolites, which can be preserved in the mudstone-dominated formations and which, as noted above, is indeed the case.

#### 7.4.2 Anoxia

Oxygen concentrations both of bottom and interstitial waters have a direct influence on the faunal composition of benthic communities (Ekdale 1988). Low O<sub>2</sub> concentrations may be indicated by a rare or absent benthic fauna together with a finely-laminated, as opposed to burrowed or mottled, sediment (Arthur et al. 1984). All of these features are characteristic of parts of the study area. For example, in some mudstone-dominated formations (e.g. Parrog Formation, parts of Tresaith and Gaerlwyd formations) ichnofossils are rare or absent (Fig. 7.2). All of these formations share a distinctive lithology comprising laminated hemipelagites in which darker organic-rich layers alternate with lighter organic-poor layers deposited during periods of high sea-level stand (see Chapter Six).

Chondrites has been used as an indicator of anoxia or low oxygen conditions by a number of authors (e.g. Bromley & Ekdale 1984; Savrdra & Bottjer 1987), although this contention may only hold for situations where Chondrites is exclusively dominant (D'Alessandro et al. 1986). More recently Vossler & Pemberton (1988) reported the presence of superabundant Chondrites and suggested that it indicated the presence of organic rich layers in dysaerobic conditions. In the study area Chondrites does not reach super-abundant or dominant levels. It is commonly most dominant in mudstone-dominated formations (e.g. Gwbert Formation, Tresaith Formation) where, as noted earlier, the only other preserved ichnofossil is Planolites.

Disrupted bedding has been observed in the Parrog Formation at Cwm-yr-eglwys (SN 016 401). The disruption of bedding may be traced laterally into undisturbed zones. It is also seen to decrease up section. Interpretation of this bedding disruption suggests that it is a product of anoxic or dysaerobic conditions present at the time of deposition and immediately afterwards.

Anoxic or low oxygen environments with relatively large amounts of organic matter would encourage the generation of gases such as methane at shallow depths within the sedimentary sequence. Overpressuring owing to compaction could possibly lead to a rupturing and an explosive escape of the organic gases up through the partially lithified sediment. The passage of these escaping gases would thus cause blocks to be disturbed in situ at the margins of the escape structure whereas those in the centre would be completely destroyed. The Parrog Formation is a mudstone-dominated formation. Muds may be extremely cohesive and therefore blocks of mud are likely to be rotated in situ and show little internal evidence, within the block, of any disturbance.

### 7.4.3 Water Depth

Seilacher's (1958, 1964, 1967, 1978) ichnofacies concept focusses on bathymetry as the primary feature governing ichnofossil distribution although it is probably not depth per se that controls the distribution but rather other factors depth-related factors, such as temperature, wave and/or current energy (Ekdale et al. 1984). As previously noted (Fig. 7.2) the ichnofaunal assemblages in the sandstone-dominated formations change both in abundance and diversity through the succession. Expressed as percentages of the total numbers of ichnogenera recorded from the entire coastal succession, the abundances rise through the formations from the Newport Sands Formation (18.75%), Poppit Sands Formation (25.0%), Llangranog Formation (50.0%), Allt Goch Formation (25.0%), Grogal Formation (12.5%) and finally the Aberystwyth Grits Formation (100.0%). The rise is not a consistent one, with some formations (e.g. Allt Goch Formation, Grogal Formation) having lower abundances than might be predicted.

The period of deposition of the Welsh Ordovician-Silurian succession was one of fluctuating sea-levels (Fig. 7.1). It is, therefore, feasible that depth, both actual and relative, was a control on the ichnofossil distribution. Actual depths, however, are very difficult to determine since there is a marked absence of specific bathymetric ranges for individual ichnofossils in the literature.

The absence of any wave-generated structures, together with the presence of graptolite-rich hemipelagic mudstone layers intercalated with turbiditic sandstones (sensu Bouma 1962) suggests that all of the formations in the study area (excepting the Parrog Formation) were deposited below storm-wave base. The area can thus be broadly classified as "deep-marine".

When the stratigraphic positions of the major sandstone lobes are compared with the suggested sea-level curves for the Welsh Basin it can be seen that there is a great deal of

apparent correlation. Sandstone depositional lobes were initiated during periods of transgression and regression, with many of them reaching a maximum extent during low sea-level stands. Thus the sandstones would have been deposited in similar water depths relative to each other. The mudstone-dominated formations are more likely to represent the deepest water depositional environments as they were, for the most part, deposited during periods of highest sea-level stands.

#### 7.4.4 Global Extinction Events

The succession in the area of study includes the Ordovician-Silurian boundary. According to Raup & Sepkowski (1982), a statistically-significant global extinction event occurred at this point in geological time. Indeed this event may be recorded in the southern Welsh Basin succession where there is a dramatic reduction in ichnogeneric diversity (from 8 to 1) across the Ordovician-Silurian boundary.

Both Seilacher (1974, 1977) and Frey & Seilacher (1980) proposed that ichnofossil diversity increased through geological time. It was suggested that shallow-marine ichnogeneric levels remained relatively static whereas those for the deep sea increased dramatically through the Phanerozoic. Given that most organisms evolved in shallow-marine environments such ideas imply increasing migration to the deep sea with time.

Analysis of thirty-three studies embracing Cambrian-Tertiary ichnofossil assemblages from all parts of the world has been undertaken to test the hypothesis (Seilacher 1974; Frey & Seilacher 1980) of increasing ichnofossil diversity through time. Only large scale studies were chosen as it was considered that these would most accurately reflect the true changes in numbers and diversity. A first plot (Fig. 7.3) uses only total numbers of ichnospecies to enable comparisons to be drawn with the work of Seilacher (1974). The resultant distribution is poor and a regression line fitted to the data

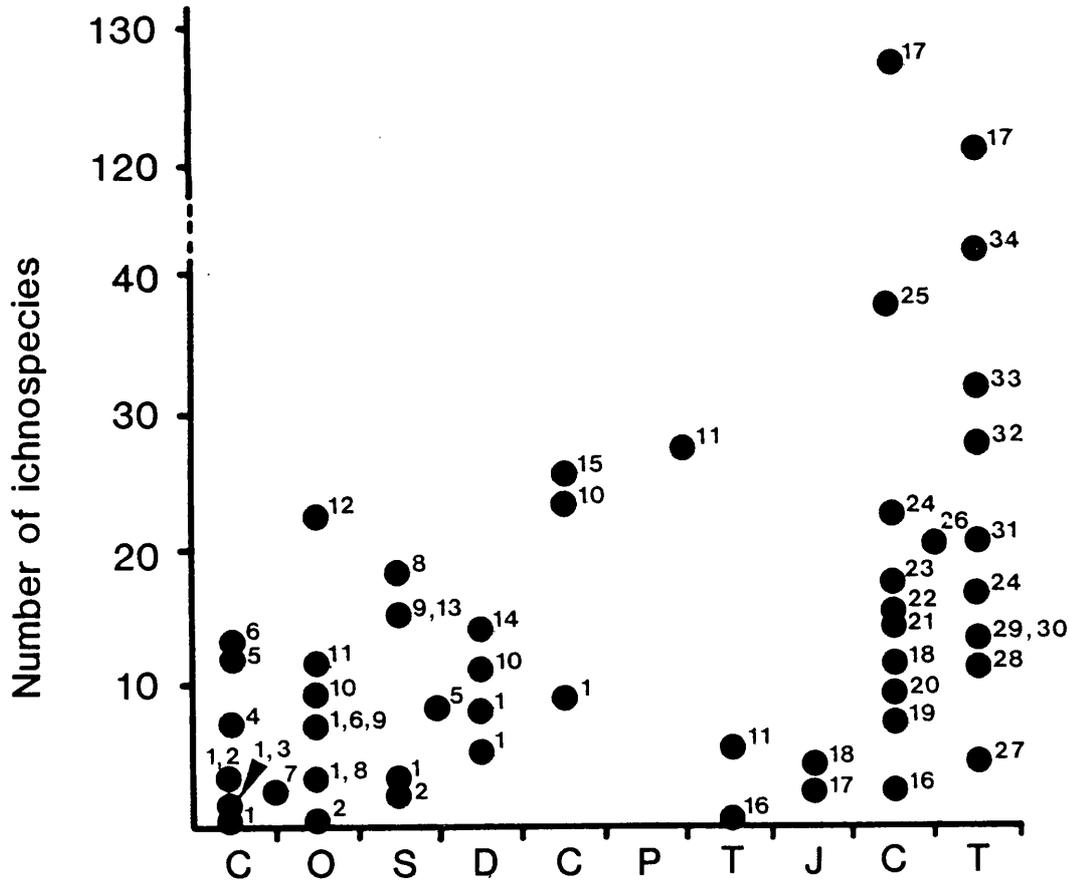


Figure 7.3 Stratigraphic distribution of deep-sea and flysch ichnospecies through geological time. Information used in this figure is as follows: 1. Cambrian-Carboniferous flysch, worldwide (Seilacher 1974); 2. Cambrian-Silurian flysch, North Greenland (Pickerill *et al.* 1982); 3. Cambrian flysch, New South Wales (Webby 1984); 4. Cambrian flysch, (Aceñolaza & Durand 1973); 5. Cambrian-Devonian flysch, Argentina (Aceñolaza 1978); 6. Cambrian-Ordovician flysch, Wales and Ireland (Crimes 1970); 7. Cambrian-Ordovician flysch, Nova Scotia (Pickerill & Keppie 1981); 8. Ordovician-Silurian flysch, Wales (present study); 9. Ordovician-Silurian flysch, Scotland (Benton 1982a); 10. Ordovician-Carboniferous flysch, East Germany (Benton 1982b); 11. Ordovician-Triassic flysch, China (Yang 1986); 12. Ordovician flysch, New Brunswick (Pickerill 1980); 13. Silurian flysch, New Brunswick (Pickerill 1981); 14. Devonian flysch, (Chamberlain 1979); 15. Carboniferous flysch, Oklahoma (Chamberlain 1971); 16. Triassic-Cretaceous flysch, northern India (Malik & Shah 1982); 17. Jurassic-Tertiary flysch, Polish Carpathians (Książkiewicz 1977); 18. Jurassic-Cretaceous flysch, Tibet (Yang *et al.* 1982); 19. Cretaceous flysch, California (Kern & Warne 1974); 20. Cretaceous flysch, Austria (Vogeltanz & Stradner 1971); 21. Cretaceous flysch, Japan (Tanaka & Sumi 1981); 22. Cretaceous flysch, Japan (Tanaka 1971); 23. Cretaceous flysch, Polish Carpathians (Radwanski 1978); 24. Cretaceous-Tertiary flysch, Austria (Kern 1978); 25. Cretaceous flysch, Alaska (McCann & Pickerill 1988); 26. Cretaceous-Tertiary flysch, Venezuela (Macsoy 1967); 27. Tertiary flysch, Venezuela (Zapata-Oviedo 1979); 28. Tertiary flysch, New Zealand (Gregory 1969); 29. Tertiary flysch, Poland (Pińkowski & Westwalewicz-Mogilska 1986); 30. Tertiary flysch, Italy (D'Alessandro 1980); 31. Tertiary flysch, Spain (Crimes 1973); 32. Tertiary flysch, Poland (Roniewicz & Pińkowski 1977); 33. Tertiary flysch, Spain (Crimes 1977); 34. Cretaceous-Tertiary flysch, Switzerland (Crimes *et al.* 1981).

gives a value of only 0.39. The line equation is  $Y = 1.27x + 3.83$ . Discounting the data point of Książkiewicz (1977, point 17; since he spent over 40 years collecting specimens and as such has an extremely large data base) does not greatly influence the results, giving a regression value of 0.49 and a line equation of  $Y = 0.67 + 6.19$ .

A second plot involves recording ichnogenera rather than ichnospecies (Fig. 7.4). In Seilacher's original work he had personally identified and speciated all the samples. Using other peoples work it is difficult to have the same degree of precision. Furthermore, given that the criteria used for ichnospeciation can be at best problematic, it is more reasonable to construct the figure at the ichnogeneric level. The resultant diagram (Fig. 7.4) also shows a poor distribution with a regression line value of 0.41 and a line equation of  $Y = 0.65 + 6.18$ . Omitting the data of Książkiewicz (1977, data point 17) merely altered the regression line value to 0.40 and the line equation to  $Y = 0.47 + 7.00$ , figures still outside the boundary of statistical acceptability.

It can easily be seen that while there is an increase in total numbers of deep-sea ichnofossil genera through time the increase is neither as dramatic nor as consistent as originally suggested by Seilacher (1974). In the revised plots of Frey & Seilacher (1980) there is a less consistent rise in numbers of deep-sea ichnospecies though the total numbers are still well below those observed by other authors. This is particularly true when examining ichnofaunal assemblages from the Palaeozoic, where diversities are more comparable with those of the Mesozoic than has been suggested (McCann & Pickerill 1988). There are, however, problems with the data set as used herein. These include the lack of studies from the Permian and Triassic periods and an overdependence on work from the Mesozoic.

A logical next step from the above analysis was to take only those ichnogenera found in the deep sea and subdivide them into two groups, namely, those which are predominantly found

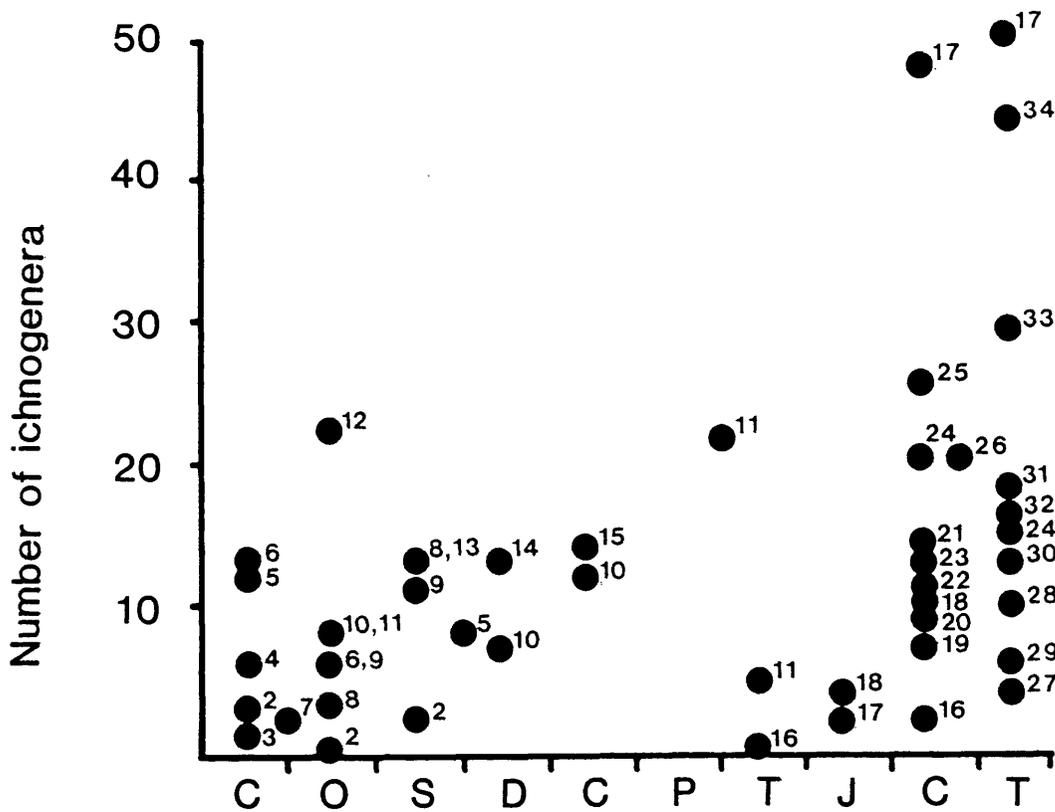


Figure 7.4 Stratigraphic distribution of deep-sea and flysch ichnogenera through geological time. See Figure 7.2 for details of information used.

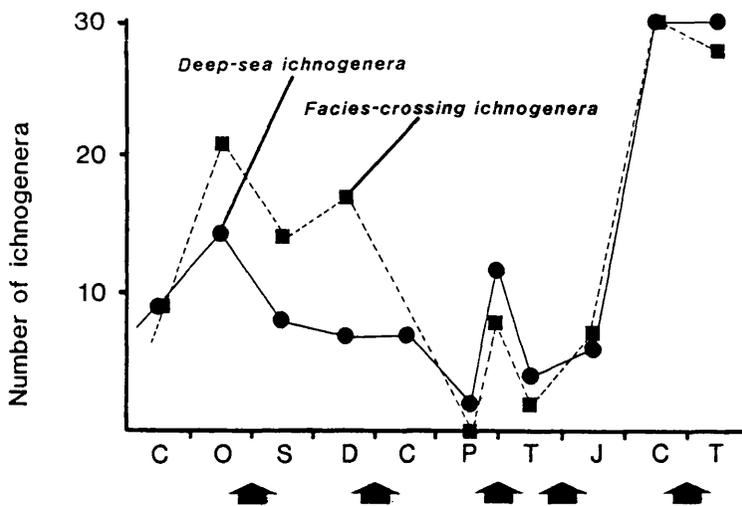


Figure 7.5 Relative distribution of both deep-sea and facies-crossing ichnogenera through geological time (See Note One). The arrows at the base represent the main marine extinctions as recorded by Raup & Sepkowski (1982).

in the deep sea (See Note One) (e.g. Belorhapse, Cosmorhapse and Helminthoida) and those which are to be found in a wide range of depositional environments (e.g. Gordia, Helminthopsis and Palaeophycus). The fact that some ichnogenera have been observed to vary in their behaviour (see Bottjer et al. 1988) was also taken into account. The resultant graph is a series of peaks and troughs wherein both the trend of the facies-crossing ichnogenera and that of the deep-sea ichnogenera closely follow each other. Troughs are taken to represent periods of marine extinction when the total numbers of ichnogenera fell. It can be seen that the pattern of the major marine extinctions (Raup & Sepkowski 1982) closely parallels the distribution pattern of troughs (Fig. 7.5). This would suggest that ichnofossil distributions in the deep-sea quite clearly reflect major changes in marine abundance and diversity as recorded in the marine fossil record. The decrease in diversity at the Ordovician-Silurian boundary (Fig. 7.5) is thus considered a reflection of the broader global situation.

## 7.5 Discussion and Conclusions

The pattern of distribution of ichnofossils in the western Welsh Basin succession is controlled by a number of variables which act either singly or, more likely, in combination. It is, therefore, difficult, if not erroneous, to suggest any single variable as being the main control on the distribution. For the area under discussion, the presence or absence of sandstone or siltstone beds (i.e. lithology) would appear to be the major controlling factor. This can be directly related to the depositional environment in which the organism may have lived, given that this determines the dominant lithology. An important adjunct to this is the quality of outcrop and the numbers of collecting sites. Fürsich (1975), in a study in the English Jurassic, noted that diversity was controlled, to an extent, by the amount of section available for examination.

It should, however, be noted that in pelagic areas of the

deep sea only infaunal burrows, emplaced below the sediment-water interface, are preserved in the fossil record (Ekdale & Berger 1978; Berger *et al.* 1979). Such infaunal forms would include both Chondrites and Planolites, the two most common ichnogenera found in the succession. Thus the apparent differences in ichnofaunal assemblage between the formations may be a function of the type of sediment (i.e. toponomy) in which the organism lived rather than any real changes in the depositional environment, and conditions within the basin. Ekdale (1988) noted that the alternation between different ichnocoenoses may be mistakenly interpreted as resulting from large scale variations in water depth.

There are, however, still some discrepancies which are not accounted for by toponomy. These are differences between formations which, superficially at least, are identical and yet their assemblages are very different. The low levels of bioturbation in some of the mudstone-dominated formations is most rationally explained by the presence of anoxic or dysaerobic bottom conditions. The fact that the low diversity mud-dominated formations tend to be the ones that contain evidence of low oxygen conditions, either by the presence of laminated hemipelagites or of areas of disrupted bedding as a result of methanogenesis, would appear to support this assertion.

For sandstone-dominated formations the causal factors are more problematic. Given that all of the formations were deposited in similar water depths (on the basis of the current sea-level curves), then one could expect to find similar ichnofaunal assemblages present in each of them. This, however, is problematic since lithologically identical facies associations from, for example, the Newport Sands or Poppit Sands formations contain far fewer ichnofossils than similar facies associations in the Aberystwyth Grits Formation. In situations such as these it is probable that other factors such as substrate assume importance. Ekdale (1988) noted that substrate variations can complicate reconstructions of palaeo-bathymetry and palaeo-oxygenation. Indeed, substrate variations can produce radically different

ichnocoenoses (e.g. Pemberton & Frey 1984). There is, however, no firm evidence that there were any differences in substrate consistency between the various sandstone formations. Indeed, the majority of the preserved ichnogenera are graphoglyptids and as such exist in broadly similar substrates.

As noted earlier, the percentages of ichnofossils are more common in the sandstone-dominated formations. Furthermore between these formations they also vary, with the emphasis being on slow increases through time from the Newport Sands Formation through to the end of the Llangranog Formation, followed by a further rise from the Allt Goch Formation through to the Aberystwyth Grits Formation. These rises may be related to the increasing colonisation of the deep-marine environment, as postulated by Seilacher (1974) and Frey & Seilacher (1980), but it is probably impossible to ascertain this with any degree of certainty.

The fall in total numbers of ichnogenera at the end Ashgill (Llangranog-Gaerglwyd formations) may be interpreted as a consequence of the major marine global extinction which occurred at the end of the Ordovician. The decrease in numbers, however, is emphasised by the transition from a sandstone-dominated formation to a mudstone-dominated formation and also from an aerobic environment to a dysaerobic one. Thus the decrease may be due to a combination of taphonomy, anoxia and possible global extinction. It is only a true result of extinction, however, if the ichnotaxa that produced the relevant trace disappear and do not reappear again. Given that ichnofossils are behavioral patterns, and as such may be produced by a variety of morphologically different and distinct organisms, it is impossible to definitively state whether extinction is the causal factor. Despite this, however, it is interesting to speculate that the fall in total numbers of ichnogenera may have been enhanced by the decrease in numbers of possible trace-making organisms as a result of the global extinction event.

The distribution of the ichnofaunal assemblages in the Welsh coastal succession is important also in terms of what can be deduced about the predictive models. The testing of Seilacher's stratigraphic model suggests that it should not be used indiscriminately. While it may be useful as a starting point when discussing ichnofossil distributions, caution should be exercised when using it in a predictive or causative fashion.

It is virtually impossible to isolate any single mechanism as being the only controlling factor for the observed distribution of ichnofossils in the study area. Ultimately it is best to consider that all of the controlling factors contributed to a greater or lesser degree. This is particularly important when applying the findings of this study to another area which may have very different environmental conditions.

**Note One:**

(a). Deep-sea (or predominately deep-sea) ichnogenera - Acanthorhappe, Belorhappe, Circulichnis, Cochlichnus, Cosmorhappe, Desmograpton, Fucusopsis, Fustiglyphus, Glockerichnus, Gyrophyllites, Halymenidium, Helicolithus, Helicorhappe, Helminthoida, Laevicyclus, Lophoctenium, Lorenzina, Megagrapton, Nereites, Oldhamia, Paleodictyon, Paleomeandron, Protopaleodictyon, Spirophycus, Spirorhappe, Strobilorhappe, Sublorenzina, Subphyllochora, Taprhelminthopsis, Unarites, Yakutatia.

(b). Facies-crossing ichnogenera - Alcyonidiopsis, Arenicolites, Bifasciculus, Buthotrepsis, Chondrites, Diplichnites, Diplocraterion, Gordia, Granularia, Gyrochorte, Helminthopsis, Mamillichnis, Muensteria, Neonereites,

Ophiomorpha, Palaeophycus, Pelecypodichnus, Phycodes,  
Planolites, Rhizocorallium, Sabularia, Saerichnites,  
Scalarituba, Scolicia, Skolithos, Taenidium, Thalassinoides,  
Zoophycos.

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CHAPTER EIGHT - A NEREITES ICHNOFACIES FROM THE  
ORDOVICIAN-SILURIAN SUCCESSION OF THE WELSH DEPOSITIONAL  
BASIN

**Abstract**

The sediments of the Ordovician-Silurian succession of west Wales are interpreted as having been deposited as a sheet sand system with associated pelagic and hemipelagic mudstones within a marginal marine basin. The strata contain a diverse trace fossil assemblage consisting of fifteen ichnogenera, namely: Chondrites, Circulichnis, Cochlichnus, Cosmorhaphe, Desmograption, Gordia, Helminthoida, Helminthopsis, Neonereites, Nereites, Paleodictyon, Palaeophycus, Planolites, Protopaleodictyon and Spirophycus. The ichnofaunal assemblage is typical of the Nereites ichnofacies.

Twenty three ichnospecies are described, one of which (Spirophycus bicornis) is noted for the first time from the Palaeozoic. Sheet sands contain the most abundant and diverse ichnoassemblages, a reflection of favorable environmental conditions for inhabitation by benthic organisms. Oxidic turbiditic mudstones contain just two ichnospecies, namely Chondrites ichnosp. and Planolites montanus both of which are common. Anoxic hemipelagites contain the same two ichnospecies albeit in greatly reduced numbers, while the hemipelagites deposited in a distal shelf environment contain the least abundant and diverse ichnoassemblage, comprising just a single specimen of Planolites montanus.

**8.1 Introduction**

The Aberystwyth Grits Formation, part of the southern Welsh Basin, has long been recognised as having an abundant and

diverse ichnofaunal assemblage (e.g. Keeping 1878, 1881). The ichnofaunal assemblage, in common with other taxonomically diverse deep-water ichnocoenoses elsewhere in the world, has not been systematically described. The present paper rectifies this by documenting the ichnofaunal assemblage of the Aberystwyth Grits Formation together with assemblages from thirteen stratigraphically older formations which outcrop along the coast from Dinas Head in the south to north of Aberystwyth. The succession in the study area is a continuous one, with a resolved stratigraphy that includes the Ordovician-Silurian boundary. Ichnofossil assemblages are reasonably common throughout the sequence.

## 8.2 Location and stratigraphy

The study area is situated on the west coast of Wales and forms part of the Lower Palaeozoic Welsh depositional basin (Bassett 1984; Woodcock 1984; Siveter et al. 1989). The area comprises a succession of fourteen formations which range in age from Llandeilo (Nemagraptus gracilis Biozone) to upper Llandovery (Monograptus turriculatus Biozone) and consistently young to the north (McCann and Pickering 1989).

The formations may be divided into two types based on the dominant lithology: (i) Sandstone-dominated formations (e.g. Newport Sands Formation, Aberystwyth Grits Formation) are deposited from high-concentration turbidity current flows. They may be classified according to the facies scheme of Pickering et al. (1986) as gravelly muds (Facies Class A), sandstones (Facies Class B) and sand-mud couplets (Facies Class C). (ii) Mudstone-dominated formations (e.g. Gwbert Formation, Tresaith Formation) comprise thin silts, muddy silts and silt-mud couplets (Facies Class D) which are laterally variable. The mudstone-dominated formations show a wide range of features consistent with deposition from dilute turbidity currents. The turbiditic beds are frequently overlain by bioturbated pelagic muds (Facies Class E) with laminated hemipelagites being locally abundant (e.g. Gaerglwyd Formation)

The distribution of the dominant lithologies appears to be determined both by eustatic sea-level changes and tectonic influences (see Chapter Four). For the strata of the study area, two major periods of transgression have been recognised in the Caradoc and the early Llandovery, respectively (McKerrow 1979; Leggett et al. 1981). These are separated by a eustatic regression in the late Ashgill which is related to a major glacial event in Gondwana (Berry and Boucot 1973; Brenchley and Newell 1984). Forzey and Cocks (1986) noted that the lithological and faunal changes in part of Wales to the south of the study area may be related to the global eustatic curve and suggested that it was not necessary to invoke a tectonic influence. All of the sandstone-dominated formations (with minor exceptions - see Chapter Four) may be related to these eustatic sea-level changes.

### 8.3 Depositional environment

For beds to be recognised as turbidites they must retain the sedimentary structures formed by the action of the turbidity current. This normally requires deposition below storm-wave base. The present study suggests that all of the sediments in the area, except those of the Parrog Formation, were deposited below storm-wave base and are possibly of relatively deeper water origin. The reasons for this interpretation are:

- a) the absence of any wave-generated structures except in the Parrog Formation, where a single occurrence of bi-directional cross bedding was noted.
- b) the presence of graptolite-rich hemipelagic mudstone layers intercalated with turbiditic sandstones. Graptolites are relatively rare in shallow water shelf deposits.
- c) the sandstones are commonly "classical" turbidites (sensu Bouma 1962).
- d) many of the sedimentary facies of Pickering et al. (1986)

and by extension those of Stow and Piper (1984) can be recognised, albeit occasionally in altered forms.

e) the absence of major slump sequences except in the Llangranog area suggests that the depositional environment was not located near a slope, as does the sheet-like aspect of many of the turbiditic sandstones.

#### 8.4 Previous Work

Much of the more recent ichnological work on the Lower Palaeozoic sediments of Wales has tended to concentrate on shallow water environments and trilobite-related ichnofossils such as Cruziana and Rusophycus (e.g. Crimes, 1968, 1969, 1970). Earlier work tended to concentrate more on deep-sea ichnoassemblages.

Murchison (1839) was the first to recognise ichnofossils from Wales. He referenced and figured four ichnospecies - Nereites cambrensis, N. sedgwickii, Myrianites macleayii (= Nereites macleayii) and Nemertites ollivantii (= ?Dictyodora) - in his classic study of the Silurian System. Keeping (1878), in a paper on the Aberystwyth area, recognised the same four ichnospecies as Murchison. He also noted that "these marvellous surface contortions and other markings....are convex on the under surface" (Keeping, 1878, his emphasis). In a paper on the geology of central Wales published in 1881, Keeping increased the number of ichnospecies to seven(?) by including Buthotrephis (= ?Chondrites), Retiofucus extensus (= ?Paleodictyon), Palaeochorda tardifurcata and "worm trails". Finally, Keeping (1882) described and figured eight ichnospecies including Buthotrephis major, B. minor (= ?Chondrites), Palaeochorda tardifurcata (= ?Helminthopsis), Retiophycus extensus (= ?Paleodictyon), Palaeophycus striatus, Myrianites lapworthii (= ?Nereites), Nematolites edwardsii and N. dendroidium (= unrecognisable "ichnogenera").

In a paper on the geology of the Llangranog area Hendricks

(1926) noted the presence of "worm trails and molluscan tracks". Price (1953) noted the occurrence of organic markings on the undersurfaces of sandstone beds in the Aberystwyth Grits. Smith (1956) and subsequently Wood and Smith (1959) noted and figured the presence of Paleodictyon. Other unnamed and unfigured structures were briefly described and appear to refer to ?Gordia and ?Planolites. Anketell (1963) noted and figured a number of ichnofossils from the Llangranog area including "gastropod trails" (= ?Nereites), and intriguingly "arthropod tracks" (= ?Diplicnites), and "echinoid traces" (= ?Astericites). Crimes (1970), in a paper demonstrating the palaeoenvironmental importance of ichnofossils, noted four ichnogenera (Nereites, Dictyodora, Paleodictyon, and Chondrites) from the Aberystwyth Grits. None of the ichnofossils, however, were described systematically nor was any indication given as to their relative frequency of occurrence. Crimes and Crossley (1980) published the most recent account of ichnofossils in the region in their examination of the two forms of Paleodictyon (Paleodictyon (Glenodictyon) and Paleodictyon (Squamodictyon)) which are found in the Aberystwyth area.

### 8.5 Systematic Ichnology

In the following section ichnofossils are considered in alphabetical order rather than any morphological or behavioral groupings. Preservational terminology follows that adopted by Häntzschel (1975) and Ekdale et al. (1984), while the internal structure of sandstone beds uses the terminology of Bouma (1962). Associated ichnofossils are those ichnofossils which are preserved on the same bedding plane as the ichnofossil under discussion.

Ichnogenus Chondrites Von Sternberg, 1833

Type ichnospecies. Fucoides antiquus Brogniart, 1828 by subsequent designation of Miller (1889).

Diagnosis. Dendritic, smooth walled, regular but asymmetrically ramifying burrow systems that normally do not interpenetrate or interconnect. Branch angles and tunnel diameters are essentially constant (after Häntzschel 1975; Pemberton and Frey 1984).

Systematic discussion. The type ichnospecies of Chondrites was designated by Miller (1889) as Fucoides antiquus and subsequent designations of Fucoides targionii (e.g. Osgood 1970; Chamberlain 1977) are invalid by rule of priority (Fillion and Pickerill 1984). Häntzschel (1975) listed Fucoides lycopodioides as the type but this is erroneous since it is not a specimen of Chondrites but rather is the type of Caulerpites Von Sternberg, 1833 (Chamberlain 1977).

The morphology of Chondrites is diverse and many ichnospecies, some of which are questionable (Crimes et al. 1981), can be found in the literature (Osgood 1970; Chamberlain 1971a). Many of the proposed ichnospecies were erected due to variations in size, preservation and mode and angle of branching (Hakes 1976). Simpson (1957), however, doubts the validity and usefulness of ichnospecies recognition in Chondrites suggesting that differences may be a result of behavioural changes effected by changing levels of nutrients in the sediment. The current situation with more than 170 ichnospecies in existence, and presumably many of these being either junior synonyms of the first few ichnospecies named by Brongniart or ontogenetic variants (Książkiewicz 1977), suggests that the ichnogenus is in severe need of systematic revision (Chamberlain 1977; Crimes et al. 1981).

Behavioral/Environmental interpretation. The origin of Chondrites has been discussed by a number of authors, notably Wanner (1949), Simpson (1957) and Osgood (1970). Chondrites may be considered as feeding or dwelling burrows or possibly both (Książkiewicz 1977). Tauber (1949) suggested a sessile, filter-feeding annelid as the possible producer, while others have suggested siphunculids (unsegmented worms), polychaetes (Simpson 1970; Hill 1971; Tanaka 1971), sea pens (Bradley

1981), small arthropods (Ekdale 1977) or unknown tentacle-bearing organisms (Taylor 1967). Recent polychaetes construct similar burrows (Bandel 1967; Pemberton and Frey 1984).

The question of whether the burrows are passively infilled or penecontemporaneous with burrow construction remains debatable (Frey 1970). Both Simpson (1957) and Osgood (1970) figured burrow reconstructions demonstrating that the Chondrites burrow system originated at a point, or points, in the sediment above which the more pinnate distal ends of the burrow system are found. Ferguson (1965) suggested that the vacuum created on the withdrawal of the proboscis from the burrow would enable detrital sediment to be sucked in. However, several ichnospecies have been reported which contain faecal pellets (e.g. Macsotay 1967) and thus, according to Frey (1970), one would have to assume that the debris surrounding the central burrow opening consisted almost exclusively of faecal pellets. Some ichnospecies having a laminated infilling have been reported (Frey 1970; Książkiewicz 1977). These would suggest that sediment was introduced periodically into open burrows or else that the tunnels may have been retrusively filled by the producing organism (Seilacher 1955; Książkiewicz 1977). Osgood (1970) has experimentally shown that clays could passively fill up to 87% of an open burrow system - the final 13% being accounted for by compaction and diagenesis.

Colour changes between the lithologies of the burrow fill and the host sediment may be due either to the action of body excretions (Griggs et al. 1969) or enzymes (Osgood 1970). The sharp contact frequently observed between the burrow filling and the tunnel wall may suggest that the walls were coated with mucus which would have also inhibited burrow collapse (Osgood 1970). Frey (1970), noting the lack of pyritised burrow lining, has suggested that the organism did not line the burrow with appreciable quantities of organic matter. The remarkable fact that the burrows rarely interpenetrate is interpreted as being due to the use of chemo-receptors by the producing organism enabling it to

achieve maximum efficiency by avoiding areas already worked (Osgood 1970; Sellwood 1970).

Examples of Chondrites in Chalk have been interpreted by Ekdale and Bromley (1983) as being the product of a deep-burrowing organism. Virtually all descriptions in both terrigenous and carbonate sediments show that the burrow was emplaced well below the sediment-water interface (Bromley and Ekdale 1984). They suggest that the presence of Chondrites indicates low O<sub>2</sub> levels in the interstitial waters within the sediment at the site of burrow emplacement. This, however, may only hold for situations where it is exclusively dominant (D'Alessandro et al. 1986). In areas where it occurs with other traces it is generally positioned deepest in the sediment. This would suggest that the producing organisms may have intentionally burrowed beneath the redox horizon to reach an organic-rich layer whereupon the burrow branched horizontally (Ekdale 1980).

Chondrites is considered to be an opportunistic form colonising the sediment soon after deposition (D'Alessandro et al. 1986; Vossler and Pemberton 1988). The Chondrites-producing organism may also have fed on faecal material. This is particularly true of smaller Chondrites forms as they came in later (Kern and Warne 1974). In areas where other burrows, for example, Thalassinoides are reworked, Chondrites penetrates to far greater depths in the burrow fill than in the surrounding sediment (Kennedy 1970).

Occurrence. The ichnogenus is a facies-crossing one (Bottjer 1982) and has been reported from a wide variety of environments. Bromley and Ekdale (1984) have suggested that oxygen-poor conditions influence its distribution far more than bathymetry or sediment type. Chondrites ranges in age from Cambrian (Singh and Rai 1983) to Eocene (Crimes et al. 1981).

Chondrites ichnospp. indet.

Pl. 8.1, Fig. a

Material. Numerous recordings in the field.

Description. Exichnially preserved, dendritic branching structures oriented parallel and sub-parallel to bedding. The mudstone is both structureless and parallel laminated. Specimens range in thickness from 0.5-6.0 cm. Branch diameters are constant within each burrow system while the branch angles vary between 30 and 50 degrees. Burrows are distinctly lined with the infill being of a darker mudstone. The system of branching is both dichotomous and pinnate. Burrow length is variable, with maximum length being 6.7 cms.

Remarks. Specimens of Chondrites are commonly divided into ichnospecies on the basis of angle of branching and burrow diameter. The systematics of the ichnogenus, however, are in need of taxonomic review and hence it is considered more appropriate to refer the specimens to Chondrites ichnospp. Planolites montanus is the only associated ichnofossil.

Distribution. Parrog Formation, Newport Sands Formation, Poppit Sands Formation, Gwbert Formation, Tresaith Formation, Llangranog Formation, Gaerglwyd Formation, Allt Goch Formation, Cefn Cwrt Formation, Llangraig Formation, Grogal Formation and Aberystwyth Grits Formation.

#### Ichnogenus Circulichnis Vialov, 1971

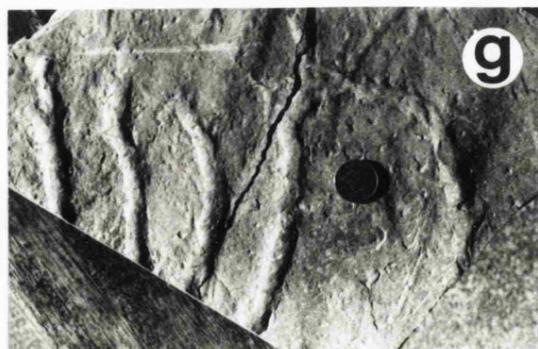
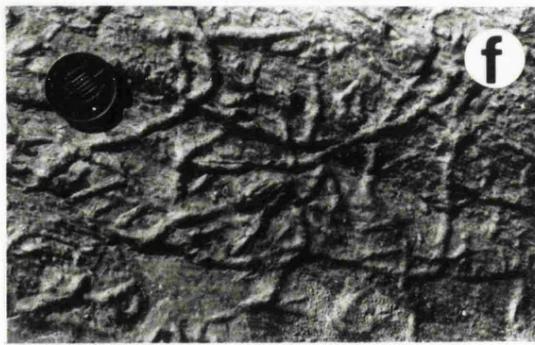
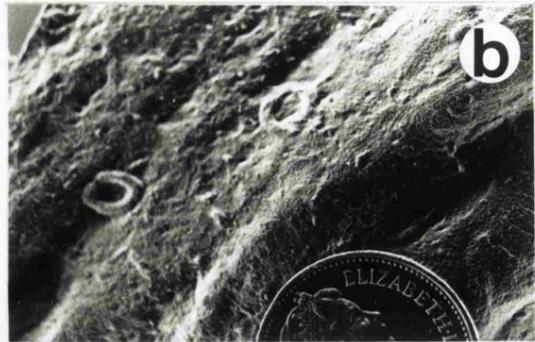
Type ichnospecies. Circulichnis montanus Vialov, 1971 by original designation.

Diagnosis. Round- or oval-shaped trace which may be in the form of a vertical cylinder within the sediment (emended from Vialov 1971, trans. litt.).

Systematic discussion. Until 1981 Circulichnis had not been formally recognised from the western hemisphere (Pickerill and Keppie 1981). Similar traces had been figured (e.g. Książkiewicz 1977) and described as "scribing traces". Pickerill and Keppie (1981) suggest that these are in fact

PLATE 8.1

- a. Chondrites ichnospp. exichnially preserved within mudstones of the Tresaith Formation (Ashgill), Penbryn Beach (SN 296534), x 0.5.
- b. Circulichnis montanus preserved on a lower bedding surface in convex hyporelief, Llangranog Formation (Ashgill), Llangranog (SN 310542), NMW89.12G.5, x 1.5.
- c. Cochlicnus anguineus preserved on a lower bedding surface in convex hyporelief, Aberystwyth Grits Formation (Upper Llandoverly), Aberystwyth (SN 583827), NMW89.12G.3, x 1.45.
- d. Cosmorhaphe sinuosa preserved on a lower bedding surface in convex hyporelief, Aberystwyth Grits Formation (Upper Llandoverly), Clarach (SN 586835), NMW89.12G.2, x 0.35.
- e. Desmograpton fuchsi preserved on a lower bedding surface in convex hyporelief, Aberystwyth Grits Formation (Upper Llandoverly), Aberystwyth, (SN 583827), NMW89.12G.1, x 1.4.
- f. Gordia marina preserved on a lower bedding surface in convex hyporelief, Aberystwyth Grits Formation (Upper Llandoverly), Gilfach-yr-Halen (SN 434613), NMW89.12G.51, x 0.4.
- g. Helminthoida ichnospp. nov. A preserved on a lower bedding surface in convex hyporelief, Aberystwyth Grits Formation (Upper Llandoverly), Aberaeron (SN 449626), NMW89.12G.6, x 0.3.
- h. Helminthopsis abeli preserved on a lower bedding surface in convex hyporelief, Aberystwyth Grits Formation (Upper Llandoverly), Gilfach-yr-Halen (SN 434613), NMW89.12G.7, x 0.9.



specimens of Circulichnis. Książkiewicz (1977), however, notes that the scribing trace comprises a raised outer ring and a series of lower inner rings. This would appear to differ fundamentally from the original description of Vialov (1971, see above) although it is morphologically closer to the ichnogenus Laevicyclus as described by Piénkowski and Westwalewicz-Mogilska (1986) from the Upper Eocene-Lower Miocene Podhale Flysch of Poland. Their specimens are both bigger and smaller than that of Książkiewicz (1977) and they also show evidence of the central burrow which is diagnostic of Laevicyclus (Häntzschel 1975). Other figured and unnamed traces which may be included in Circulichnis include those of Häntzschel (1975, fig. 44, 2a, p. W71), Książkiewicz (1958, Table III, fig. 2), Kitchell et al. (1978, fig. 3, 17, p. 174), Kitchell and Clark (1979, Plate 4, fig. 2, p. 1054) and Byers (1982, fig. 6, 17, p. 239). The "circling trace fossils" of Crimes et al. (1981) may also be included since they are unornamented and are therefore not the same as the scribing trace of Książkiewicz (1977).

The original diagnosis of Circulichnis is confusing since it is clear in the rest of the text that it is in fact a burrow and not a trail which is being described (Fillion and Pickerill 1984). The same authors also noticed evidence of burrow collapse and that the burrow fill was lithologically the same as the surrounding matrix.

Behavioral/Environmental interpretation. The method of production of Circulichnis is obscure (Crimes et al. 1981). Vialov (1971) suggested that since no exit was observed that the producer presumably moved back up or down into the sediment once the "whorl" was completed. Pickerill and Keppie (1981) figured specimens of Helminthopsis ichnosp. which were of similar dimensions to associated specimens of Circulichnis. They have suggested that the producer of both ichnofossils was possibly an annelid, with Circulichnis representing a behavioral variant. A similar case could also be suggested for the specimen figured by Książkiewicz (1957) where Cosmorhappe and Circulichnis are also of similar dimensions. It is therefore probable that Circulichnis had a

number of possible producers.

Recent Lebensspuren with morphologies resembling Circulichnis have been reported from the Venezuela Basin of the Caribbean Sea (Young et al. 1985). The two features, one a circular groove the other a circular ridge were suggested as being produced by the oweniid polychaete Myriochele. Members of this genus have been reported by Hartman and Fauchald (1971) from slope to abyssal depths.

Due to the paucity of formal recordings it is not wise to use Circulichnis as a specific environmental indicator (Pickerill and Keppie 1981). It is possibly a eurybathic form being reported from shallow marine carbonates in Quebec (Fillion and Pickerill 1984) although most of the recordings are from deeper water environments.

Occurrence. The ichnogenus ranges in age from Tremadoc (Fillion and Pickerill 1984) to Recent (Kitchell et al. 1978; Kitchell and Clark 1979).

Circulichnis montanus Vialov, 1971

Pl. 8.1, Fig. b

Diagnosis. As for ichnogenus.

Material. Two specimens collected (NMW89.12G.5) and one field observation.

Description. Smooth, ellipsoidal traces preserved in convex hyporelief on the soles of 57.0–123.0 mm thick, parallel- to cross-laminated, fine- to medium-grained sandstone. Burrows are cylindrical, unlined and 1.0–1.3 mm in diameter. Infill is identical to that of the host rock. The diameter of the long axis of the traces ranges from 4.2–5.7 mm while the shorter axis ranges from 1.2–3.0 mm.

Remarks. Three of the present specimens are found on the base of a sandstone bed which also contains groove casts.

This would suggest that Circulichnis is a post-depositional burrow. A fragment of a Circulichnis montanus is also found on the same bed. While this fragment is morphologically similar to Gordia arcuata it does differ in a number of respects. The fragment is isolate unlike G. arcuata which is commonly gregarious. It is also of similar dimensions to the specimens of Circulichnis montanus on the same surface. The fragment is thus regarded as a partial specimen of C. montanus rather than G. arcuata. Associated ichnofossils include Desmograpton fuchsi, Helminthopsis ichnosp., Palaeophycus tubularis, Palaeophycus ichnosp., and Paleodictyon (Glenodictyon) imperfectum.

Distribution. Llangranog and Aberystwyth Grits formations.

#### Ichnogenus Cochlichnus Hitchcock, 1858

Type ichnospecies. Cochlichnus anguineus Hitchcock, 1858 by original designation.

Diagnosis. Regularly winding, smooth horizontal burrows or trails resembling the sine curve (after Hakes 1976; Pemberton and Frey 1984).

Systematic discussion. The synonymies and nomenclatural history of Cochlichnus have been discussed by a number of authors including Michelau (1955), Nowak (1970), Häntzschel (1975) and Hakes (1976). Cochlichnus is often confused with Belorhappe Fuchs, 1895 another winding ichnofossil (e.g. Michelau 1955). Belorhappe, however, is characterised by angular bends rather than the sinusoidal curves of Cochlichnus (Hakes 1976). Häntzschel (1975) considered smooth sinusoidal forms as Cochlichnus and maintained Belorhappe as a separate ichnogenus.

Behavioral/Environmental interpretation. Although originally classed by Häntzschel (1975) as a flysch ichnofossil, Cochlichnus is now considered to be a eurybathic form occurring in a wide variety of environments ranging from

river mouth deposits (Yang et al. 1987) and river floodplains (Fordyce 1980) to shallow marine environments (Banks 1970; Webby 1970; Pemberton and Frey 1984; Crimes and Anderson 1985) and deep sea turbidites (Aceñolaza 1978; Piénkowski and Westwalewicz-Mogilska 1986). Despite its wide-ranging occurrence Cochlichnus has been used as a facies indicator in the Upper Carboniferous cyclothem of West Germany where Michelau (1955) interpreted it as representing an environment intermediate between the brackish water Planolites ophthalmoides facies and the fresh water pelecypod facies.

Cochlichnus was probably produced by vermiform organisms (Pemberton and Frey 1984) or annelids lacking well developed parapodia (Hakes 1976). Fordyce (1980) noted that Cochlichnus resembles fossil nematode trails as figured by Moussa (1969, 1970). Moussa (1969) observed that the sinusoidal trails were only produced when a nematode moved over a mud surface covered with a thin film of water no thicker than its body. This, however, would restrict the occurrence of Cochlichnus to a narrow range of environments which, as previously noted, is not the case. Fordyce (1980) suggested that the size of larger trace-making organisms would have allowed trace formation in relatively deeper waters.

Occurrence. Cochlichnus ranges in age from late Precambrian (Aceñolaza and Durand 1973; Palić et al. 1979) to Oligocene (Häntzschel 1975).

Cochlichnus anguineus Hitchcock, 1858

Pl. 8.1, Fig. c

Diagnosis. As for ichnogenus.

Material. Two specimens collected (NMW89.12G.3, 4).

Description. Unbranched, smooth, gently sinusoidal burrow preserved in convex hyporelief on the base of a 10.0-40.0 mm thick, Tcde or Tde, fine- to medium-grained sandstone. The

burrow is 0.3-2.0 mm in diameter and up to 54.00 mm in length. The amplitude of the sine curve is 1.2-2.5 mm while the wavelength is 7.0-13.0 mm.

Remarks. The sinusoidal winding is considered diagnostic of the ichnogenus and serves to distinguish it from the more irregular winding and meandering burrows of Helminthopsis. Fillion (1984), in his discussion of the ichnogenus, concluded that it was preferable to consider it as monospecific. There are no associated ichnofossils.

Distribution. Aberystwyth Grits Formation.

#### Ichnogenus Cosmorhapse Fuchs, 1895

Type ichnospecies. Cosmorhapse sinuosa Azpeitia-Moros, 1933 by subsequent designation of Häntzschel (1975).

Diagnosis. Unbranched, composite, meandering burrows consisting of wide first order meanders with superimposed second order meanders (after Książkiewicz 1977; Seilacher 1977).

Systematic discussion. The nomenclatural history of Cosmorhapse is confused. The ichnogenus was originally illustrated by Fuchs (1895, pl. 6, fig. 1) who gave neither a diagnosis nor a species name (Książkiewicz 1977). Subsequently Häntzschel (1975) chose Cosmorhapse sinuosa Azpeitia-Moros, 1933 as the type specimen. While this has made Cosmorhapse a valid ichnogenus it has also altered the original meaning since the holotype of Cosmorhapse sinuosa (Azpeitia-Moros 1933, fig. 24b) undulates both less regularly and with a smaller amplitude than the specimen originally figured by Fuchs (1895). The composite meanders, however, are diagnostic of the ichnogenus and serve to differentiate it from the highly organised meanders of Helminthoida and the loop-like traces of Phycosiphon.

Behavioral/Environmental interpretation. According to

Książkiewicz (1977) Cosmorhaphe is possibly both a pre-depositional and a post-depositional form. It is typical of deep water environments (Książkiewicz 1958; Vogeltanz 1971; Zapata Oviedo 1979; D'Alessandro 1980) with similar traces being observed in the Kermadec Trench at depths of 4,735m (Bourne and Heezen 1965). Rodriguez and Gutschick (1970) reported Cosmorhaphe from shallow marine carbonates of late Devonian - early Mississippian age but their figured specimens (p. 432, pl. 6a) are more akin to ?Phycosiphon or ?Helminthopsis since they lack the composite meanders of Cosmorhaphe. Modern day Lebensspuren resembling Cosmorhaphe are interpreted as infaunal burrow systems positioned just below a veneer of sediment (Ekdale 1980).

Originally interpreted as the impression of gastropod spawn strings (Fuchs 1895; Macsotay 1967), Cosmorhaphe is now recognised as a typical feeding burrow, although some are possibly locomotion trails (Książkiewicz 1977) produced by worm-shaped organisms such as Acorn worms (Webby 1969) or polychaetes.

Occurrence. The ichnogenus ranges in age from ?Ordovician to Carboniferous (Volk 1964; Aceñolaza 1978; Baldis and Aceñolaza 1978) and Cretaceous to Recent (Ekdale 1980; Crimes et al. 1981).

Cosmorhaphe sinuosa Azpeitia-Moros, 1933

Pl. 8.1, Fig. d

Diagnosis. Widely spaced first order meanders with superimposed second-order meanders in which the wavelength is less than the amplitude. Occasional short cuts may connect successive turns of the ichnofossil (Seilacher 1977).

Material. One specimen (NMW89.12G.2).

Description. Smooth, unbranched, composite meanders preserved in convex hyporelief on the sole of a 45.0 mm thick, Tbce, fine- to medium-grained sandstone. Burrow

diameter ranges between 4.0-5.0 mm with a slight thickening at the apices of the first-order meanders. The first-order meanders range from 110.0-130.0 mm in wavelength and from 43.0-56.0 mm in amplitude. They are superimposed by smaller second-order meanders which range from 30.0-45.0 mm in wavelength and from 7.0-13.0 mm in amplitude.

Remarks. Cosmorhaphe sinuosa may be distinguished from the related C. fuchsi by the fact that both the first and second order meanders are constricted at the base. Cosmorhaphe gracilis has been interpreted as a juvenile form of C. sinuosa (Książkiewicz 1977). Gordia marina is the only associated ichnofossil.

Distribution. Aberystwyth Grits Formation.

#### Ichnogenus Desmograpton Fuchs, 1895

1895 Desmograpton Fuchs, pp. 394-395, pl. 5, figs. 1-2, 4-6.

1967 Pseudodesmograpton Macsotay p. 36

non 1971 Baroccoichnites Vialov p. 88

1977 Desmograpton (Fuchs); Książkiewicz p. 181

non 1978 Desmograpton Fuchs; Radwanski, p. 56, plate 2, fig. 1-3

Type ichnospecies. Desmograpton fuchsi Książkiewicz, 1977 by subsequent designation.

Description. Biramous burrows with lateral appendages in which either the transverse or longitudinal elements are pronounced (Seilacher 1977).

Discussion. Detailed description, synonymy and interpretation of this distinctive ichnogenus was recently published by McCann (1989). This information is summarised in chapter nine.

Desmograpton fuchsi Książkiewicz, 1977

Pl. 8.1, Fig. e

- 1895 Desmograpton Fuchs, pp. 394-395, pl. 5, figs. 1-2, 4-6.
- 1954 Desmograpton Fuchs, Seilacher, p. 218, fig. 14.
- 1959 Desmograpton Fuchs, Seilacher, p. 1069, fig. 15.
- non 1967 Pseudodesmograpton ichthyiformis Macsotay, p. 36, pl. 6, fig. 20.
- ?1967 Helminthopsis Heer; Macsotay, pl. 6, fig. 20.
- 1977 Desmograpton ichthyforme (Macsotay); Seilacher, pp. 311-312, fig. 7d.
- 1977 Desmograpton inversum Seilacher, p. 312, fig. 7f.
- 1977 Desmograpton fuchsi Książkiewicz, pp. 182-183, pl. 29, fig. 5, text fig. 43.

Diagnosis. Hypichnial, roughly straight and parallel burrows with some transversal links (Książkiewicz 1977).

Material. One specimen (NMW 89.12G.1).

Description. Smooth burrow system preserved in convex hyporelief on the soles of 0.5-5.7 cm thick, Tcde or Tde, fine- to medium-grained sandstones. The overall length of the structures is from 15.0-37.0 mm, with burrow diameter ranging between 3.0-8.0 mm. The specimen shows pronounced medial thickening of the parallel to sub-parallel burrow elements. The transverse links are arranged alternately and range from 5.0-12.0 mm in length and from 1.5-2.3 mm in diameter.

Remarks. Associated ichnofossils include Circulichnis montanus, Helminthopsis ichnosp., Palaeophycus tubularis, Palaeophycus ichnosp. and Paleodictyon (Glenodictyon) imperfectum.

Distribution. Aberystwyth Grits Formation.

Type ichnospecies. Gordia marina Emmons, 1844 by monotypy.

Diagnosis. Smooth, loosely winding and meandering traces which cross themselves extensively (after Książkiewicz 1977).

Systematic discussion. Gordia was established by Emmons (1844) for an irregularly looping burrow with numerous crossovers from the Silurian of southeastern Maine. The original description of Emmons was rather scanty and the above diagnosis was based instead on the figure provided by Emmons (1844, pl. 14, fig. 24) and redrawn in Książkiewicz (1977 fig. 36r). The definition given by Häntzschel (1975) does not mention the diagnostic crossovers and should therefore be disregarded. The schematic drawings (Häntzschel 1962, fig. 121.2; 1975, fig. 39, 1b) do not correspond to the figures given by Emmons (1844) and subsequently by Hall (1847). Instead the drawings depict a Helminthopsis-type trace (Książkiewicz 1977). Häntzschel (1975), however, also figured a specimen of Gordia from Fischer and Paulus (1969) which does show crossovers and is therefore accepted as a true Gordia. This tendency to cross its own burrow is a feature absent from all other meandering traces and thus sets Gordia apart (Crimes et al. 1981). Hence the lack of crossovers would preclude the identification of a trace as Gordia.

Gordia has been synonymised with a number of ichnogenera erected subsequent to its original designation. Palaeochorda, established by M'Coy (1848, 1851) for traces from the Lake District of England, contained two ichnospecies separated on a size basis alone. Since the topotype material has a full size range, they are best considered as specimens of Gordia. Nicholson (1873, 1978) synonymised both Gordia and Palaeochorda with Nemertites ollivantii Murchison, 1839. The specimens of Murchison, however, are basal walls of the ichnogenus Dictyodora and not complete traces. Hence the synonymy is considered inappropriate (Benton 1982).

Behavioral/Environmental interpretation. The ichnogenus is

produced by worms or worm-like organisms (Aceñolaza 1978) or small snails (Bandel 1967). Abel (1935) pictured some worm trails on North Sea tidal flats which were similar to Gordia, while Nereis produces similar traces. It is essentially a eurybathic form occurring in a range of environments from littoral and shallow marine (Bandel 1967; Yang 1984) to deep-water turbidites (Benton 1982).

Occurrence. Gordia ranges in age from ?Precambrian to Cainozoic (Häntzschel 1975).

Gordia marina Emmons, 1844

Pl. 8.1, Fig. f

Diagnosis. Smooth, unbranched, horizontal trails or burrows, of uniform diameter throughout their length, winding but not meandering and with numerous crossovers (Fillion 1984).

Material. Five specimens (NMW89.12G.47, 48, 49, 50, 51) collected and several occurrences noted in the field.

Description. Smooth, unbranched burrows, 2.0-4.0 mm in diameter, preserved in convex hyporelief on the soles of parallel and cross-laminated, 18.0-49.0 mm thick, fine- to medium-grained sandstones. Burrow diameter is constant throughout length. The burrows wind irregularly and cross themselves up to three times.

Remarks. Gordia marina may be distinguished from G. arcuata Książkiewicz, 1977 since in the latter only the apical arcuate bends are preserved. Fillion (1984) suggested that Gordia molassica Heer, 1865 should be considered as a junior synonym of Gordia marina since their respective courses are similar and size ranges are unknown. Associated ichnofossils include Cosmorhaphé sinuosa, Palaeophycus tubularis, Palaeophycus ichnosp. and Paleodictyon (Glenodictyon) imperfectum.

Distribution. Llangranog and Aberystwyth Grits formations.

Ichnogenus Helminthoida Schafhautl, 1851

Type ichnospecies. Helminthoida labyrinthica Heer, 1865

Diagnosis. Simple unbranched meandering trails or burrows which loop back and forth in closely-spaced parallel to subparallel paths on bedding planes (Häntzschel 1975; Kern 1978).

Systematic discussion. Helminthoida is a variable ichnogenus (Książkiewicz 1977). Burrows are commonly parallel and closely spaced (Crimes and Anderson 1985) although they can be extremely variable (Häntzschel 1975). Meander size and regularity are probably controlled by the use of stimuli by the Helminthoida-producing organism, for example homostrophy, thigmotaxis and phototaxis (Häntzschel 1975).

To accommodate some of the observed variety Książkiewicz (1970) divided the ichnogenus into two groups with the Helminthoida labyrinthica group being further subdivided into a number of subspecies. It is, however, doubtful whether such divisions are necessary given that some of the forms are merely preservational variants. Seilacher (1977) separated Helminthoida crassa Schafhautl, 1851 from H. labyrinthica, as he considered the former to be an open tunnel, and erected Helminthorhapse as a new ichnogenus to emphasise the difference. Crimes et al. (1981) and subsequently Crimes and Anderson (1985) discussed the subdivision and considered it unnecessary. It should also be noted that part of Seilacher's (1977) reasoning was the fact that Helminthoida labyrinthica occurred only in shales. McCann and Pickerill (1988) reported Helminthoida labyrinthica from the Cretaceous of Alaska preserved in hyporelief on the base of sandstone beds.

Behavioral/Environmental interpretation. Previous interpretations of Helminthoida have suggested algae, spawn strings, feeding traces of gastropods (Häntzschel 1975) and body impressions of worms (Gregory 1969) as probable producers. It is now regarded as the internal grazing trails

of worms or worm-like organisms (Książkiewicz 1977). The similarities between Helminthoida and the faecal casts of enteropneusts from the Recent has also been noted (Cullen 1967).

Occurrence. The ichnogenus ranges in age from Cambrian (Crimes and Anderson 1985) to Recent (Fujioka et al. 1987) and is considered to be typical of deep-water flysch environments (Crimes 1977). It should, however, be noted that Crimes and Anderson (1985) reported some specimens from shallow marine environments which suggests that the form originated there and later migrated to the deeper ocean.

Helminthoida ichnosp. nov. A

Pl. 8.1, Fig. g

Material. One specimen (NMW89.12G.6).

Description. Simple meandering trace preserved in convex hyporelief on the sole of a 14.0 mm thick, cross-laminated siltstone to fine-grained sandstone bed. Burrow fill is identical to that of the host sediment. The trace is 10.0mm wide and has a granulose textured surface which comprises a series of bumps without any ordered arrangement. The meanders have an amplitude of 23.0–67.0 mm and a minimum wavelength of 130 mm. It is not possible to determine the true wavelength due to erosion of part of the specimen. Sectioning of the specimen reveals that it is, in fact, a cast of either a burrow or a trail.

Remarks. The systematic meandering pattern suggests that this ichnofossil belongs to the ichnogenus Helminthoida Schafhautl, 1851 rather than the less regular winding burrows and trails of the ichnogenus Helminthopsis Heer, 1877. The specimen conforms well with other figured and reported specimens of Helminthoida in the literature (c.f. Książkiewicz 1977). The granulose ornament, however, has not been described before in this ichnogenus. The texture may have been produced as a result of the stuffing, by the

burrowing organism, of pellets of sediment into a burrow. There are no associated ichnofossils.

Distribution. Aberystwyth Grits Formation.

Ichnogenus Helminthopsis Heer, 1877

Type ichnospecies. Helminthopsis abeli Książkiewicz, 1977 by subsequent designation of Książkiewicz (1977).

Diagnosis. Irregularly winding burrows and trails with a tendency to meander (after Fillion and Pickerill 1984).

Systematic discussion. In the original description of Helminthopsis, Heer (1877) established three ichnospecies but designated no type species (Fillion and Pickerill 1984). Subsequently a number of authors (e.g. Ulrich 1904; Andrews 1955) designated Helminthopsis magna as the type ichnospecies. This, however, is a bilobate specimen and examination of the type material shows that it belongs to the ichnogenus Scolicia and thus the synonymy is considered to be invalid (McCann and Pickerill 1988). Książkiewicz (1977) has since proposed a specimen illustrated by Abel (1935, fig. 261B), which conforms better to the original diagnosis of Heer, as the type ichnospecies under the name Helminthopsis abeli.

The ichnogenus is divided into ichnospecies according to the burrow diameter, the type of winding and the presence or absence of surface ornament (Książkiewicz 1970; Pickerill 1981). Reassessment, however, of these criteria is required since, in practice, they are extremely variable (Pickerill 1981). As this would necessitate complete re-examination of all type, co-type and topotype material, and is thus outside the scope of this study, the criteria outlined above will be used herein for identification of ichnospecies.

Behavioral/Environmental interpretation. The ichnogenus is a post-depositional trace produced by worm-like organisms,

presumably polychaetes and possibly priapulid annelids (Książkiewicz 1977). Carey (1978) has suggested that specimens from Permian barrier beach sediments in the Sydney Basin may be the remnants of gastropod feeding trails. Crimes et al. (1981) also reported a specimen stuffed with faecal pellets for part of its length. Helminthopsis is essentially a eurybathic form although it is most frequently reported from deep-sea environments.

Occurrence. It ranges in age from Eocambrian (Aceñolaza and Durand 1973) to Eocene (Crimes et al. 1981).

Helminthopsis abeli Książkiewicz, 1977

Pl. 8.1, Fig. h

Diagnosis. Loosely winding Helminthopsis with a tendency to meandering. Meanders are irregular and variable in shape (Książkiewicz 1977).

Material One specimen (NMW89.12G.7) collected and rare recordings in the field.

Description. Smooth, loosely winding and meandering burrows preserved in convex hyporelief on the soles of 20.0–25.0 mm thick, normally graded and parallel-laminated, fine-grained sandstone and siltstone. Burrow diameter, which is constant throughout, ranges from 2.3–4.0 mm and length is up to 127 mm. Infill is identical to that of the host sediment.

Remarks. Helminthopsis abeli may be distinguished from other ichnospecies of Helminthopsis by its tendency to meander (Książkiewicz 1977). Helminthopsis ichnosp. is the only associated ichnofossil.

Distribution. Aberystwyth Grits Formation.

Helminthopsis hieroglyphica Heer in Maillard, 1887

Pl. 8.2, Fig. a

Diagnosis. Irregularly winding *Helminthopsis* where the course of the trace is alternately winding and straight (Książkiewicz 1977).

Material. Two specimens (NMW89.12G.8, 9) and rare recordings in the field.

Description. Smooth, alternately straight and winding burrow preserved in convex hyporelief on the sole of a 10.0–65.0 mm thick normally graded, fine- to medium-grained sandstone and siltstones. Burrow diameter is 3.0–9.0 mm and constant throughout the length of up to 230.0 mm. Infill is identical to that of the host sediment.

Remarks. *Helminthopsis hieroglyphica* differs from *H. tenuis* in its smaller diameter (Książkiewicz 1977). *Helminthopsis* ichnosp. is the only associated ichnofossil.

Distribution. Aberystwyth Grits Formation.

*Helminthopsis granulata* Książkiewicz, 1968

Pl. 8.2, Fig. b

Diagnosis. Irregular winding and loosely meandering *Helminthopsis* where the surface is covered with bumps and ridges arranged parallel to the axis of the burrow (Książkiewicz 1977).

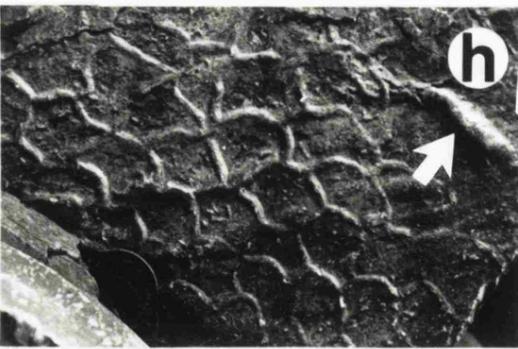
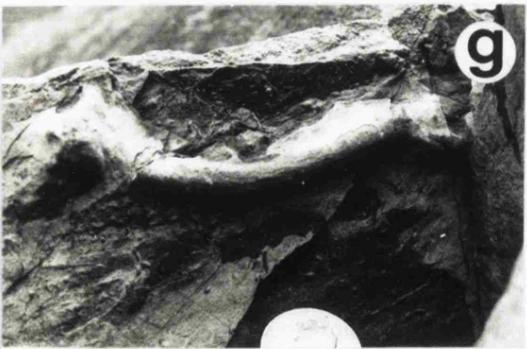
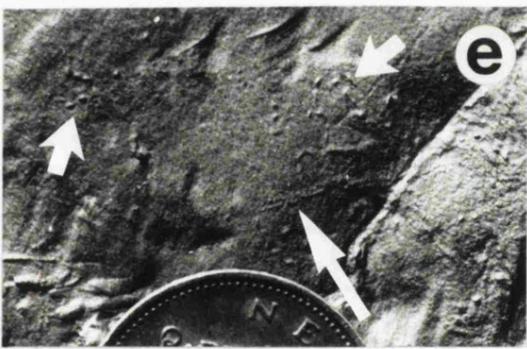
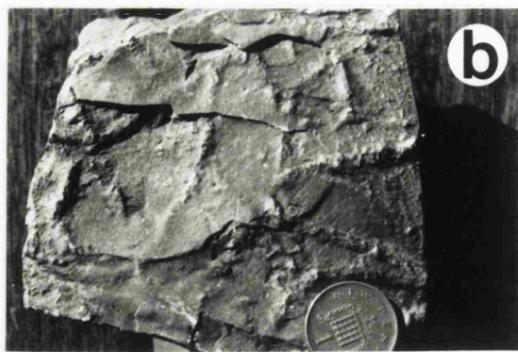
Material. One specimen (NMW89.12G.12).

Description. Loosely winding burrow preserved in convex hyporelief on the sole of a 50.0 mm thick, Tbcde, fine- to medium-grained sandstone. Burrow diameter is 3.0 mm and is constant throughout the length of 37.0 mms. The surface of the burrow has a granulated, wart-like texture.

Remarks. *Helminthopsis granulata* may be distinguished from other ichnospecies of *Helminthopsis* by its granulated surface ornamentation. There are no associated ichnofossils.

PLATE 8.2

- a. Helminthopsis hieroglyphica preserved on a lower bedding surface in convex hyporelief, Aberystwyth Grits Formation (Upper Llandoverly), Aberystwyth (SN 583827), x 0.5.
- b. Helminthopsis granulata preserved on a lower bedding surface in convex hyporelief, Aberystwyth Grits Formation (Upper Llandoverly), New Quay (SN 385604), NMW89.12G.12, x 0.65.
- c. Helminthopsis cf. irregularis preserved on a lower bedding surface in convex hyporelief, Aberystwyth Grits Formation (Upper Llandoverly), New Quay (SN 385604), NMW89.12G.10, x 0.45.
- d. Helminthopsis ichnosp. preserved on a lower bedding surface in convex hyporelief, Aberystwyth Grits Formation (Upper Llandoverly), Aberaeron (SN 449626), NMW89.12G.16, x 0.55.
- e. Neonereites uniserialis (short arrows) and Nereites cf. macleayi (long arrow) preserved on an upper bedding surface in convex epirelief, Aberystwyth Grits Formation (Upper Llandoverly), New Quay (SN 385604), NMW89.12G.23, x 2.05.
- f. Palaeophycus anulatus preserved on a lower bedding surface in convex hyporelief, Aberystwyth Grits Formation (Upper Llandoverly), New Quay (SN 385604), NMW89.12G.32, x 0.7.
- g. Palaeophycus tubularis preserved on a lower bedding surface in convex hyporelief, Llangranog Formation (Ashgill), Llangranog (SN 310542), x 0.7.
- h. Palaeophycus ichnosp. (arrowed) and Paleodictyon (Glenodictyon) imperfectum preserved on a lower bedding surface in convex hyporelief, Aberystwyth Grits Formation (Upper Llandoverly), Aberystwyth (SN 583827), x 0.5.



Distribution. Aberystwyth Grits Formation.

Helminthopsis cf. irregularis (Schafhautl, 1851)

Pl. 8.2, Fig. c

Material. Two specimens (NMW89.12G.10, 11) and rare recordings in the field.

Description. Gently winding and meandering burrows preserved in convex hyporelief on the sole of a 25.0-30.0 mm thick, Tcde or Tde, fine- to medium-grained sandstone or siltstone. Burrow diameter is 3.0-5.0 mm and is constant throughout the length of up to 160.0 mm. The surface of the burrow is transversely annulated with up to 4 annulae per millimetre. The traces have an irregular meander.

Remarks. The ichnospecies is discussed in detail by Książkiewicz (1977) whose specimens consist mainly of trails with portions of burrow fill still present. The faint transverse segmentation is observed in the trails rather than on the surface of the burrow. The present specimen, however, comprises a full burrow with transverse annulae on the surface of the burrow and as such is only tentatively referred to Helminthopsis irregularis. There are no associated ichnofossils.

Distribution. Aberystwyth Grits Formation

Helminthopsis ichnosp.

Pl. 8.2, Fig. d

Material. Ten specimens (NMW89.12G.13, 14, 15, 16, 17, 18, 19, 20, 21, 22) collected and numerous field occurrences.

Description. Smooth, straight to winding burrows and trails, variably preserved in concave epirelief and convex hyporelief on the surfaces of 9.0-64.0 mm thick, structureless, parallel- or cross-laminated mudstone to medium-grained

sandstone. Burrow fill is identical to that of the host sediment while diameter ranges from 0.4-7.0 mm. The maximum burrow length is 45.0 mm.

Remarks. Due to poor preservation these specimens were only identified to the ichnogeneric level. Associated ichnofossils include Circulichnis montanus, Desmograpton fuchsi, Helminthopsis abeli, H. hieroglyphica, Palaeophycus anulatus, Palaeophycus ichnosp. and Protopaleodictyon submontanum.

Distribution. Poppit Sands Formation, Llangranog Formation, Allt Goch Formation, Aberystwyth Grits Formation.

#### Ichnogenus Neonereites Seilacher, 1960

Type ichnospecies. Neonereites biserialis Seilacher, 1960 by original designation.

Diagnosis. Irregularly winding epichnial trace consisting of smooth dimples. The corresponding hyporeliefs comprise rows of pustules (Seilacher 1960).

Systematic discussion. Without thorough systematic revision it remains uncertain whether Neonereites constitutes an ichnogenus in its own right. Some authors have regarded both Neonereites and Nereites as preservational variants of Scalarituba Weller, 1899 (Chamberlain 1971a) although Seilacher and Meischner (1965) grouped all three together as informal synonyms. Hakes (1976) and Pickerill (1981) regard all three as distinct ichnogenera.

Behavioral/Environmental interpretation. Neonereites is a facies-crossing ichnogenus found in both shallow (Crimes and Anderson 1985; Yang 1984) and deep-water environments (Crimes 1977; Hill 1981; Yang et al. 1982). It is a post-depositional trace (Crimes et al. 1981; Seilacher 1962) interpreted as a crawling trail (Häntzschel and Reineck 1968) or burrow (Häntzschel 1975) made by a deposit-feeding worm or

worm-like organism (Chamberlain 1977).

Occurrence. The trace ranges in age from Precambrian (Fedonkin 1976) to Tertiary (Häntzschel 1975).

Neonereites uniserialis Seilacher, 1960  
Pl. 8.2, Fig. e

Diagnosis. Neonereites consisting of straight to winding, uniserially arranged dimples in epirelief and nodes in hyporelief (Seilacher 1960).

Material. One specimen (NMW89.12G.23).

Description. Winding to meandering rows of small (0.3-0.6 mm) uniserially arranged knobs preserved in convex epirelief on the surface of a 15 mm thick cross-laminated siltstone. The unornamented, circular to ovoid knobs form a chain up to 14.0 mm long. Knob spacings range from 0.5-2.0 mm.

Remarks. The uniserial arrangement of the knobs in the specimen distinguish the specimen from N. biserialis. Associated ichnofossils include Nereites cf. macleayi and Helminthopsis ichnosp.

Distribution. Aberystwyth Grits Formation.

Ichnogenus Nereites MacLeay, 1839

Type ichnospecies. Nereites cambrensis Murchison, 1839 by subsequent designation of Häntzschel (1962).

Diagnosis. Irregularly meandering epichnial trace comprising a smooth to annulated narrow median furrow. This is flanked on both sides by closely spaced, spherical, ovate or pinnate lobes which are commonly striated (Crimes and Germs 1982; Hakes 1976).

Systematic discussion. The position assumed by the Nereites-producing organism determines to a large extent the morphology of the resultant trace (Hakes 1976). This variation, coupled with the variety of morphological features used for ichnospecific recognition, has led to large numbers of ichnospecies being recorded (e.g. Delgado 1910). Critical re-examination of the ichnogenus in recent years has reduced the number of ichnospecies to four (Benton 1982) although the relative importance of lobe shape and size, both of which are strongly controlled by preservational factors, still needs to be evaluated (McCann and Pickerill 1988). There also remains the fundamental problem as to the nature of the morphological relationship between Nereites and Neonereites (see Neonereites section for further details) and between both of these and Scalarituba.

Behavioral/Environmental interpretation. Nereites were formerly regarded as plants, worms or graptolites (Häntzschel 1975) but are now interpreted as the internal grazing and locomotion trails of deposit feeding infaunal organisms (Seilacher 1983). Possible producers include worms (Chamberlain 1971b; Aceñolaza and Durand 1973; Aceolaza 1978), sea pens (Bradley 1981), gastropods (Abel 1935; Macsotay 1967) or crustaceans (Häntzschel 1975). The ichnogenus is commonly found in deep-water environments and is the type ichnogenus of the Nereites ichnofacies of Seilacher (1964, 1967). It should, however, be noted that specimens of Nereites have been reported from shallow-water environments (Garcia-Ramos 1976; Crimes and Anderson 1984). and the ichnogenus should be regarded as eurybathic.

Occurrence. Nereites ranges in age from Precambrian (Crimes and Germs 1982) to Tertiary (Aceñolaza 1978).

Nereites cf. macleayi Macleay in Murchison, 1839  
Pl. 8.2, Fig. e

Material. Two specimens (NMW89.12G.23) on the same slab.

Description. Straight epichnial trails up to 12.0 mm long with a smooth median furrow, 0.4-1.0 mm wide and constant throughout, flanked on both sides by rounded lobes preserved on the top of a 15.0 mm thick, cross-laminated siltstone bed. Lobe diameters range from 0.3-0.7 mm and have a density of 8-21 per cm.

Remarks. Based on the criteria of Benton (1982) regarding lobe shape and size the specimens from west Wales may be assigned to N. macleayi. The lack of any meandering or winding of the traces, however, precludes a definite assignment of the specimens.

The specimens herein differ in two main respects from those of Benton (1982). Firstly the lobe diameters are smaller and secondly they are more closely spaced. Despite this they still conform most closely with the ichnospecific diagnosis of N. macleayi rather than any of the other three ichnospecies discussed by Benton (1982) and would therefore serve to widen the dimensional definition of the ichnospecies. Associated ichnofossils include Helminthopsis ichnosp. and Neonereites uniserialis.

Distribution. Aberystwyth Grits Formation.

#### Ichnogenus Palaeophycus Hall, 1847

Type ichnospecies. Palaeophycus tubularis Hall, 1847 by subsequent designation of Miller (1889).

Diagnosis. Lined, unbranched, rarely branched, straight to slightly curved or undulose, smooth or ornamented burrows of variable diameter. Infillings are typically structureless and of the same lithology as the host rock (Pemberton and Frey 1982; Fillion and Pickerill 1984).

Systematic discussion: An exhaustive review of the ichnogenus was undertaken by Pemberton and Frey (1982) which examined in detail the previous research, synonymies and the

problem of differentiating between Palaeophycus and the related ichnogenus Planolites Nicholson, 1873. Distinguishing the two ichnogenera had been a major problem. Many factors had been used to try to differentiate them including the presence or absence of true branching (Alpert 1975) and whether or not the lithology of the burrow fill was identical to that of the host rock (Osgood 1970). Pemberton and Frey (1982, 1984) distinguish the two primarily on the basis of wall lining and the character of burrow fill. Infills of Palaeophycus represent passive, gravity-induced sedimentation and therefore the lithology of the burrow fill is the same as that of the surrounding matrix, unless the overlying sediments are different (Fillion and Pickerill 1984) or there is concealed bed-junction preservation (Hallam 1975). Incomplete infilling results in the presence of burrow collapse features (Fillion and Pickerill 1984).

Pemberton and Frey (1982) followed Häntzschel (1975) in taking Palaeophycus tubularis, as designated by Bassler (1915), as the type. More recently Fillion and Pickerill (1984) have noted that Palaeophycus tubularis had already been designated as the type ichnospecies by Miller (1889).

Species of the ichnogenus are recognised according to the thickness of the burrow wall and the presence or absence of surface ornament (Pemberton and Frey 1982).

Behavioral/Environmental interpretation. Although Palaeophycus is typically interpreted as a dwelling structure (Pemberton and Frey 1982) it may also be a crawling or feeding trace (Chamberlain 1977, 1979; Hofmann 1979). It is, however, most definitely a burrow (Osgood 1970). The hypichnial and endichnial preservation suggests that the tracemaking organism foraged along sedimentologic interfaces (Pemberton and Frey 1984). As the organism moved along at depth the sediment was pushed aside producing tensional microfaults or differentially compacted laminations (Pemberton and Frey 1982).

Similar traces, produced by predaceous polychaetes (e.g.

Glycera), have been observed in the Recent (Osgood 1970). The branching burrow network produced by Glycera alba, G. americana and G. dibranchiata possess mucus-coated walls (Pemberton and Frey 1982). Similarly mucus-coated burrows of the nereid polychaetes Nereis virens and N. succinea vary from simple branching structures to a quite complex array with prominent vertical and horizontal components (Howard and Frey 1975). Passive infilling of these burrows has also been observed (Pemberton and Frey 1982).

While both of these genera are compatible this does not restrict the analogy to polychaetes. Gastropods (e.g. Polynices), pelecypods (Osgood 1970) and annelids (Hallam 1970) have also been suggested as possible producers. Palaeophycus is a generalised form and consequently may be produced by different organisms in different environments. The observed differences in sculpture (e.g. Hofmann 1979) would tend to support this. Palaeophycus is a eurybathic form found in both marine and non-marine environments.

Occurrence. Palaeophycus ranges in age from Precambrian (Nowlan et al. 1985) to Recent (Aceñolaza 1978).

Palaeophycus anulatus Fillion, 1984

Pl. 8.2, Fig. f

Diagnosis. Palaeophycus with distinct annulations (Fillion 1984).

Material. One specimen (NMW89.12G.32) and one noted in the field.

Description. Lined burrows preserved in convex hyporelief on the soles of 11.0–37.0 mm thick mudstone to medium-grained sandstone which is parallel-laminated and normally graded. The burrows are 4.0–5.0 mm in diameter, straight to very gently winding, and up to 55 mm in length. Burrow fill is identical to that of the host rock. The annulae are faintly marked with 2–4 per mm.

Remarks. Palaeophycus anulatus may be distinguished from other ichnospecies of Palaeophycus by its transverse annulation. This may be a result of peristaltic movement (Fillion 1984). Associated ichnofossils include Helminthopsis ichnosp., Palaeophycus tubularis and Palaeophycus ichnosp.

Distribution. Aberystwyth Grits Formation

Palaeophycus tubularis Hall, 1847

Pl. 8.2, Fig. g

Diagnosis. Smooth, unornamented Palaeophycus of variable diameter, thinly but distinctly lined (Pemberton and Frey 1984).

Material. Two specimens (NMW89.12G.33, 34) collected and numerous field occurrences.

Description. Smooth, unbranched, horizontal, straight to slightly curved burrows preserved in convex hyporelief and concave epirelief on the surfaces of 30.0-180.0 mm thick, normally graded and cross-laminated, fine- to coarse-grained sandstone with rare load and groove casts. Burrow diameter ranges from 7.0-16.0 mm and is constant throughout. Maximum length is 98.0 mm. Burrow infill is identical to that of the host rock and is structureless. The burrows are distinctly lined.

Remarks: The thinner wall lining serves to distinguish Palaeophycus tubularis from P. heberti while the lack of surface ornamentation distinguishes it from other ichnospecies of Palaeophycus. Associated ichnofossils include Circulichnis montanus, Desmograpton fuchsi, Gordia marina, Palaeophycus anulatus, Palaeophycus ichnosp. and Protopaleodictyon submontanum.

Distribution. Poppit Sands Formation, Llangranog Formation, Allt Goch Formation, Aberystwyth Grits Formation.

Palaeophycus ichnosp.

Pl. 8.2, Fig. h

Material. Two specimens (NMW89.12G.30, 31) and numerous field observations.

Description. Smooth, unbranched, unlined or thinly lined burrows preserved in convex hyporelief and concave epirelief on the surfaces of 1.5-72.0 cm thick, normally graded, parallel- and cross-laminated, coarse siltstone to granule-grade sandstone. Burrow diameter ranges from 0.3-8.0 mm and the structures are up to 90.0 mm in length.

Remarks. The lining, where present, in these specimens is always thin and this would suggest that most of these specimens probably belong to Palaeophycus tubularis. Poor preservation, however, prevents fuller identification. Associated ichnofossils include Circulichnis montanus, Desmograpton fuchsi, Gordia marina, Helminthopsis ichnosp., Palaeophycus anulatus, P. tubularis and Protopaleodictyon submontanum.

Distribution. Newport Sands Formation, Poppit Sands Formation, Llangranog Formation, Allt Goch Formation, Aberystwyth Grits Formation.

Ichnogenus Paleodictyon Meneghini in Murchison, 1850

Diagnosis. Honeycomb-like network of ridges, preserved in hyporelief, consisting of polygons which may be four- to eight-sided (Häntzschel 1975).

Systematic discussion. The systematics of Paleodictyon are in considerable disarray and in need of review. In recent years a number of papers have dealt with the problem (e.g. Książkiewicz, 1977; Seilacher, 1977; Crimes and Crossley, 1980) and an amalgamation of these ideas will be followed herein. Papers discussing earlier literature include Azpeitia-Moros (1933), Sacco (1939), Simpson (1967) and

Wanner (1949).

Vialov and Golev (1960, 1965) were the first to subdivide the ichnogenus. Four divisions were proposed, namely Paleodictyon (Paleodictyon) and Paleodictyon (Glenodictyum) as subichnogenera, and Pleurodictyon and Squamodictyon as two separate ichnogenera. It should be noted that their use of Glenodictyum was incorrect as the term Glenodictyon (see Sacco 1939) has historical priority. The original subdivisions of Vialov and Golev (1960, 1965) were based both on the thickness of the mesh ribs (which as pointed out by Książkiewicz (1977) may vary in the same specimen) and the shape and form of the mesh. Häntzschel (1962) considered both Pleurodictyon and Glenodictyon to be junior synonyms of Paleodictyon.

The confusion engendered by the various schemes has led to the proposal herein of a simplified version incorporating all of the major morphological features encountered in the ichnogenus. It is hoped that this classificatory scheme will prove workable until a major revision of the ichnogenus is undertaken. The system used herein is as follows:

a) Paleodictyon (Glenodictyon) - used for regular hexagonal-type meshes.

b) Paleodictyon (Squamodictyon) - used for petal-shaped meshes.

Both of these primary divisions are based upon mesh shape. The division retains Squamodictyon as a subichnogenus of Paleodictyon, the merit of which was previously noted by Crimes and Crossley (1980). Recognition of ichnospecies within each grouping follows the schemes of Książkiewicz (1977) and Seilacher (1977) for Glenodictyon-type traces and Seilacher (1977) and Crimes and Crossley (1980) for Squamodictyon-like traces.

c) Paleodictyon (Glenodictyon) forma pleurodictyonides - subgroup used to denote specimens with abnormal ribbing, and,

d) Paleodictyon (Glenodictyon) forma punctata - subgroup to denote specimens composed of tubercules.

These two forms may be considered to equate with Pleurodictyon Vialov and Golev, 1960 and Ramidictyon Seilacher, 1977 respectively.

Behavioral/Environmental interpretation. Many authors have discussed the origin of Paleodictyon suggesting a variety of explanations, including algal impressions, infilled mud cracks and interference ripple marks (Koriba and Miki 1939; Sacco 1939; Wanner 1949; Gregory 1969; Osgood 1970). It is now generally accepted that they were produced by organisms burrowing at a sand/mud interface. Both Glenodictyon and Squamodictyon were maintained as open tunnel systems, possibly reinforced by mucous lining, and only passively infilled by sand during or after the deposition of the overlying turbidite bed (Seilacher 1977; Crimes and Crossley 1980). Examination of Lebensspuren retrieved from modern day deep seas would tend to support this (Ekdale 1980). The pre- or post-depositional origin of the ichnofossils has also been extensively discussed (e.g. Simpson 1967; Książkiewicz 1977) with a number of authors suggesting that both are possible (Dzulynski and Sanders 1962; Książkiewicz 1970; Crimes 1973).

Paleodictyon is generally accepted as being indicative of deep water environments and it is a characteristic ichnofossil of Seilacher's Nereites ichnofacies (Kindelan 1919; Seilacher 1967; Weber 1973; Radwanski 1978; Yang et al. 1982). Crimes and Anderson (1985), however, reported it from shallow marine deposits in the Cambrian of Newfoundland and suggested that the trace-producing organism originated in a shallow marine environment and subsequently migrated into the less competitive deep-sea one.

Occurrence. The ichnofossil ranges in age from Cambrian (Crimes and Anderson 1985) to Recent (Ekdale 1980).

Paleodictyon (Glenodictyon) imperfectum Seilacher, 1977  
Pl. 8.2, Fig.h; Pl.8.3, Fig. a

Diagnosis. Large netlike structures with wide meshes of unequal size and shape (after Seilacher 1977).

Material. Four specimens (NMW89.12G.36, 37, 38, 39) and several observations in the field.

Description. Five- to seven-sided netlike-like structure preserved in convex hyporelief on the base of parallel and cross-laminated 25.0-110.0 mm thick, fine- to medium-grained sandstone beds. Strings are smooth, 1.0-3.0 mm in diameter, and have a constant thickness throughout. Individual mesh diameters range from 20.0-90.0 mm.

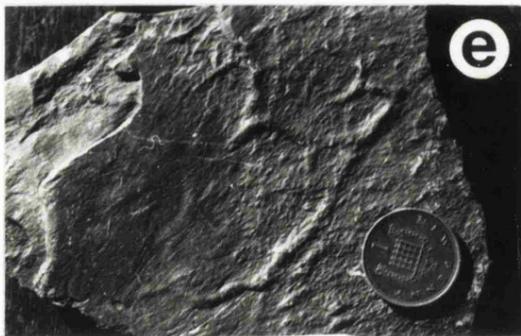
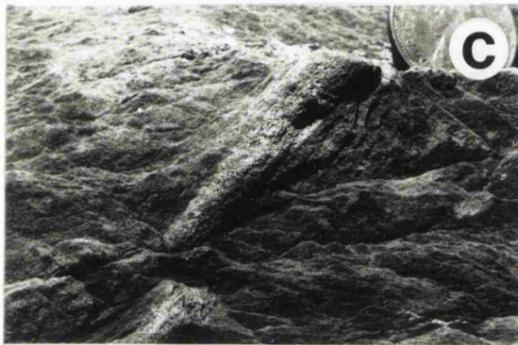
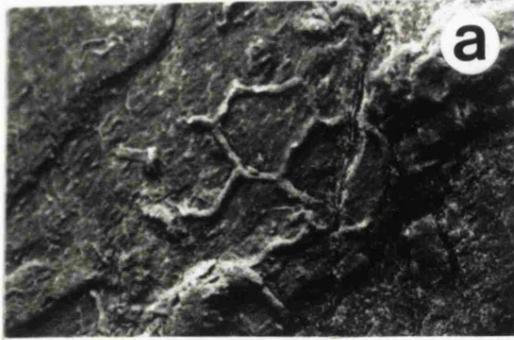
Remarks. Some of the better preserved specimens are more regular in appearance than others although all are comparable with the illustration of Seilacher (1977, Fig. 14d). One specimen is of particular interest in terms of its situation vis-a-vis the overlying sediment. The positioning of the specimen suggests that the burrow was not cast by the incoming medium and was instead either an open burrow which was filled by sediment filtering down from above or else that the burrow was backfilled by the producing organism.

In the absence of any evidence of sediment backfill structures it is herein suggested that the burrow was passively infilled by overlying sediment. The incoming turbidite bed did not erode down as far as this particular specimen and thus the entire feature has been preserved. The mudstone overlying the burrow fill is possibly part of the original overlying mudstone within which the Paleodictyon-producing organism lived. Associated ichnofossils include Gordia marina, Helminthopsis ichnosp., Palaeophycus tubularis and Palaeophycus ichnosp.

Distribution. Aberystwyth Grits Formation.

PLATE 8.3

- a. Paleodictyon (Glenodictyon) imperfectum preserved on a lower bedding surface in convex hyporelief, Aberystwyth Grits Formation (Upper Llandovery), Aberystwyth (SN 583827), x 0.8.
- b. Paleodictyon (Squamodictyon) petaloideum preserved on a lower bedding surface in convex hyporelief, Aberystwyth Grits Formation (Upper Llandovery), Aberystwyth (SN 583827), x 0.5.
- c. Planolites montanus exichnially preserved within a mudstone bed, Newport Sands Formation (Llandeilo), Carregedryw (SN 048418), NMW89.12G.35, x 0.8.
- d. Protopaleodictyon cf. incompositum preserved on a lower bedding surface in convex hyporelief, Aberystwyth Grits Formation (Upper Llandovery), Aberystwyth (SN 583827), NMW89.12G.26, x 1.75.
- e. Protopaleodictyon submontanum preserved on a lower bedding surface in convex hyporelief, Aberystwyth Grits Formation (Upper Llandovery), New Quay (SN 393596), NMW89.12G.24, x 0.65.
- f. Spirophycus bicornis preserved on a lower bedding surface in convex hyporelief, Aberystwyth Grits Formation (Upper Llandovery), Aberaeron (SN 449626), NMW89.12G.29, x 0.65.



Paleodictyon (Squamodictyon) petaloideum Seilacher, 1977

P. 8.3, Fig. b

Diagnosis. Meshes without consistent angular bends, arranged like scales or petals around a centre. The outline of the whole mesh is rounded (Seilacher 1977)

Material. Seven specimens (NMW89.12G.40, 41, 42, 43, 44, 45, 46) collected and several observed in the field.

Description. Irregular network patterns preserved in convex hyporelief on the soles of normally graded, parallel or cross-laminated, 31.0-65.0 mm thick fine- to medium-grained sandstone beds. Burrow diameters range from 1.2-3.0 mm and are constant throughout. Meshes are scale-like, petaloid and irregular in form with widths of 14.0-47.0 mm diameter and breadths of 10.0-19.0 mm.

Remarks. Paleodictyon (Squamodictyon) petaloideum may be distinguished from P.(S.) tectiforme Sacco, 1886 by its larger size. Helminthopsis ichnosp. is the only associated ichnogenus.

Distribution. Aberystwyth Grits Formation.

Ichnogenus Planolites Nicholson, 1873

Type ichnospecies. Planolites vulgaris Nicholson and Hinde, 1875 (= Palaeophycus beverleyensis Billings, 1862) by subsequent designation of Alpert (1975).

Diagnosis. Lined and unlined, rarely branched, straight to tortuous, smooth to irregular, walled, or annulated burrows which are circular to elliptical in cross-section. The burrows are of variable dimensions and configurations. The infillings are essentially structureless and differ in lithology from the host rock (Pemberton and Frey 1982; Fillion and Pickerill 1984).

Systematic discussion. The ichnogenus was originally proposed by Nicholson (1873) for certain fossilised annelid burrows, in an abstract of a paper that was never published. The missing manuscript was recently discovered and published by the University of Aberdeen with the addition of a discussion by Benton and Trewin (1978).

In his original paper Nicholson (1873) did not designate a genotype but assigned three ichnospecies (Planolites vulgaris, P. granosus, P. articulatus) to the genus, the first, Planolites vulgaris, becoming the type ichnospecies. Subsequently Nicholson and Hinde (1875) described the ichnospecies and the ichnogenus in some detail. Following the extensive review of both Planolites and Palaeophycus by Pemberton and Frey (1982) three ichnospecies, namely P. montanus, P. beverleyensis (= P. vulgaris Nicholson, 1893; = P. articulatus Nicholson, 1893) and P. annularis, were recognised. Ichnospecies are recognised on the basis of size, curvature and ornament.

The problems of differentiating Planolites from Palaeophycus have been outlined in the ichnogenic discussion on Palaeophycus. Planolites may be easily distinguished from Palaeophycus since the lithology of its burrow fill is different from that of the surrounding rock. Pemberton and Frey (1982, 1984) regarded Planolites as an unlined form but the diagnosis was emended by Fillion and Pickerill (1984) to include lined forms and therefore the presence or absence of lining can no longer be used as a valid criterion. Concealed bed junctions (c.f. Hallam 1975, fig. 4.1) may be a potential problem but according to Pemberton and Frey (1984) their presence can be correctly ascertained with careful petrologic or petrographic study of given specimens. The fill may also differ in fabric (e.g. Heinberg 1970), composition and colour (Pemberton and Frey 1982; Frey and Bromley 1985).

Colour changes observed in the burrow fill are either geochemical (Hill 1981) or biochemical in origin (Ekdale 1977). Biochemical alterations of sediment passing through the acidic microenvironment of an organism's gut may be

substantial and include changes in clay mineralogy, absorption of organic components and reduction in grain size of carbonates. These changes may also make the infill more resistant to weathering (Howell 1943).

Behavioral/Environmental interpretation. Planolites is generally attributed to the sediment-ingesting activities of deposit-feeding organisms (Banerjee et al. 1975) although it lacks the evidence of active infilling (e.g. meniscate infill of Muensteria von Sternberg, 1833) except in some rare examples (Pemberton and Frey 1984). It has also been suggested that it is partly gravity filled (Palij et al. 1979). In a behavioral pattern reminiscent of Chondrites, Planolites has been observed to reburrow numerous other traces (Frey and Chowns 1972). The reason for this is probably because the pre-existing traces contained nutritious organic residues (Frey and Seilacher 1980) and/or they represented an easier path for new burrowing activities (Pemberton and Frey, 1984).

The ichnogenus is eurybathic and is found in a wide range of depositional environments from littoral (Wright and Benton 1982; Raina et al. 1983; Singh and Rai 1983) through shallow water (Cowie and Spencer 1970; Frey 1970; Ward and Lewis 1975; Vossler and Pemberton 1988) to deep-water deposits (Ekdale 1980; Yang et al. 1982). It is, however, used as a guide fossil for the Upper Carboniferous Augenschiefer of West Germany (Häntzschel 1975) where the non-marine facies is represented by Planolites montanus Richter, (1937) and marine to brackish conditions by Planolites ophthalmoides Jessen, 1970 (Hakes 1976).

Given that the Planolites burrow is a generalised one it is highly probable that there were many different producing organisms (D'Alessandro 1980). It is mostly a post-depositional form and is rarely pre-depositional (Pińkowski and Westwalewicz-Mogilska 1986). Although there are no direct modern analogues, similar behavioral patterns have been shown by some forms, for example, Heteromastus filiformis, a deposit feeding capitellid polychaete, which

forms a branching network of interconnected feeding burrows. The eunicid polychaete, Marphysa sanguinea, and the enteropneust, Balanoglossus, both of which form branched feeding networks, are other possible analogues (Pemberton and Frey 1982).

Occurrence. Planolites ranges in age from Precambrian to Recent (Pemberton and Frey 1982).

Planolites montanus Richter, 1937

Pl. 8.3, Fig. c

Diagnosis. Relatively small Planolites with curved to contorted burrows (Pemberton and Frey 1982).

Material. One specimen (NMW89.12G.35) collected and numerous field observations.

Description. Unlined, circular to ellipsoidal, gently curved burrows preserved exichnially within 5.0-48.0 mm thick mudstone. The burrows are up to 65.0 mm long while diameters range from 1.0-3.4 mm. Infill is structureless and is lithologically different from that of the host rock.

Remarks: The specimens of Planolites montanus described herein are gently curving rather than tortuous. Their size range, however, falls within the limits of P. montanus as defined by Pemberton and Frey (1982). This also serves to distinguish them from the larger burrows of P. beverleyensis, Chondrites ichnospp. is the only associated ichnofossil.

Distribution. Parrog Formation, Newport Sands Formation, Poppit Sands Formation, Gwbert Formation, Mwnt Formation, Tresaith Formation, Llangranog Formation, Cefn Cwrt Formation, Llangraig Formation, Grogal Formation, Aberystwyth Grits Formation.

Ichnogenus Protopaleodictyon Książkiewicz, 1970

Type ichnospecies.      Protopaleodictyon incompositum  
Książkiewicz, 1970

Diagnosis. Uniramous and biramous burrows consisting of wide first order meanders and sine-shaped second order undulations. The branches usually originate at the apex of the meanders (Książkiewicz 1977; Seilacher 1977).

Systematic discussion. The ichnogenus Protopaleodictyon (= Protopalaeodictyon Książkiewicz, 1958) was first erected by Książkiewicz (1970) without either a formal diagnosis or a named type ichnospecies (Häntzschel 1962). Initially this oversight, coupled with the difference in spelling, led to some confusion (e.g. Protopalaeodictyum Nowak, 1959). Książkiewicz (1970), however, provided a diagnosis, a type ichnospecies and changed the spelling to that in current usage. A number of ichnogenera including Unarites Macsotay, 1967, Pseudopaleodictyon Pfeiffer, 1968, Spinorhaphe Pfeiffer, 1968 and Anapaleodictyon Tanaka, 1970 have been placed into synonymy with Protopaleodictyon (Häntzschel 1975; Seilacher 1977). Książkiewicz (1977) also notes that traces previously assigned by him to Palaeochorda (e.g. Książkiewicz 1961, 1970) should also be included in the ichnogenus.

According to Pickerill et al. (1982) there is still no agreement as to which characters constitute a distinct ichnospecies of Protopaleodictyon. Książkiewicz (1977) favours the regularity and spacing of the first order meanders and the size and thickness of the trace while Seilacher (1977) uses the number of branches and undulations. There is, however, enough overlap between the two schemes to ensure that a workable classificatory system, based on meander and branch pattern, exists.

Protopaleodictyon is a meandering trace with variable appendages and, as the name suggests, it can form irregular networks. However the actual transition from Protopaleodictyon to Paleodictyon remains to be documented (Seilacher 1977). Häntzschel (1975) states that it represents a combination of features from Cosmorhaphe and

Belorhaphe. Protopaleodictyon, however, can be distinguished both from Cosmorhaphe, by the presence of lateral ramifications (Książkiewicz 1970), and from Belorhaphe, by its more sinuous course, its longer appendages and the fact that its lateral branches are largely horizontal (Seilacher 1977).

Behavioral/Environmental interpretation. While Protopaleodictyon is essentially a deep-water ichnofossil (Häntzschel 1975), Benton (1982) placed it in an environment of fluctuating depth. It has also been reported from shallow marine deposits in Newfoundland (Crimes and Anderson 1985). It would therefore probably be best to regard it as eurybathic with a distinct deeper water bias. Protopaleodictyon is generally interpreted as a feeding trail made by vagile, deposit-feeding organisms (Tanaka 1971), possibly infaunal annelids (Pickerill 1981).

Occurrence. It ranges in age from Cambrian (Crimes and Anderson 1985) to Eocene (Arita 1971).

Protopaleodictyon cf. incompositum Książkiewicz, 1970  
Pl. 8.3, Fig. d

Material Three specimens (NMW89.12G.26, 27, 28).

Description. Smooth, winding burrows, 0.5-1.3 mm in diameter, preserved in convex hyporelief on the sole of a 37.0-55.0 mm thick, Tabce or Tace, fine- to medium-grained sandstone. The first order meander is not preserved. The gently winding second-order meander has a wavelength of 1.0-2.0 mm and an amplitude of 1.0-3.5 mm. These meanders are branched, frequently from the apical areas, with the branches being up to 6.0 mm in length.

Remarks: The lack of any first order meanders precludes the definite ichnospecific assignation of the two specimens. There are no associated ichnofossils.

Distribution. Llangranog and Aberystwyth Grits formations.

Protopaleodictyon submontanum Azpeitia-Moros, 1933

Pl. 8.3, Fig. e

Diagnosis. Hypichnial, irregularly winding and meandering burrows branching at various points but mostly at the apices of the meanders. They occasionally form irregular networks (Książkiewicz 1977).

Material. Two specimens (NMW89.12G.24, 25) and some rare field occurrences.

Description. Irregularly winding smooth burrows preserved in convex hyporelief on the sole of 7.0-20.0 mm thick, parallel to cross-laminated, coarse siltstone to fine-grained sandstone. The burrows are 2.0-3.0 mm in diameter with branching at various points but commonly at the apical bends. The branches are straight to gently curved.

Remarks: The ichnospecies may be distinguished from Protopaleodictyon incompositum in its irregularity, larger branches and a more pronounced net-forming tendency (Książkiewicz 1977). Associated ichnofossils include Helminthopsis ichnosp., Palaeophycus tubularis, Palaeophycus ichnosp., and Paleodictyon (Glenodictyon) imperfectum.

Distribution. Aberystwyth Grits Formation.

Ichnogenus Spirophycus Häntzschel, 1962

Type ichnospecies. Spirophycus bicornis (Heer 1876)

Diagnosis. Smooth or patterned traces which are spirally coiled at one or both ends. Both the trace and the manner of coiling are extremely variable (Książkiewicz 1977).

Systematic discussion. The ichnogenus was originally noted

by Heer (1877) who described the trace as a specimen of Muensteria von Sternberg, 1833. This name, however, was already in use for a type of retrusively filled burrow and subsequently Häntzschel (1962) renamed the ichnogenus. The type ichnospecies, Muensteria bicornis, therefore became Spirophycus bicornis. Häntzschel (1962) also placed a number of other forms into synonymy with the ichnogenus (e.g. Muensteria caprisa Heer, 1877; M. involutissime Sacco, 1888; Ceratophycus Schimper, 1879).

The ichnogenus comprises traces which are spirally coiled at one or both ends (Crimes et al. 1981). It may be preserved as casts (Seilacher 1962) and full burrows (Książkiewicz 1977). Some specimens have been reported with discoloured zones up to 2.0 cm thick around the burrow. According to Kern and Warme (1974) this suggests that the traces are faecal strings within larger burrows. They also note that the rugose surface ornament sometimes observed suggests that they are burrows stuffed with faecal pellets. Książkiewicz (1977) noted specimens lined with small foraminifera.

Behavioral/Environmental interpretation. Spirophycus is both a pre- (Seilacher 1962) and post-depositional (Książkiewicz 1977) burrow recorded almost exclusively from deep-sea turbiditic environments (Książkiewicz 1970; Kern and Warme 1974; Tanaka and Sumi 1981) although similar traces have been reported from Tertiary shallow-water environments (Cullen 1967) with a dubious occurrence recorded from the Devonian of China (Yang et al. 1984). The trace is produced by polychaetes, acorn worms (Heezen and Hollister 1971) or gastropods (Macsoy 1967).

Occurrence. Spirophycus ranges in age from Silurian (this study) to Tertiary (Häntzschel 1975).

Spirophycus bicornis (Heer 1876)

Pl. 8.3, Fig. f

Diagnosis. Hypichnial smooth burrow or trail curved like a

crozier at one or both ends. There is considerable variation in both the manner of coiling and the shape beyond the coiled portion (Książkiewicz 1977).

Material. One specimen (NMW89.12G.29).

Description. Smooth unbranched trace, spirally coiled at one end with the other opening out into an irregular meander, preserved in convex hyporelief on the base of a 23.0 mm thick, fine- to medium-grained Tcde sandstone. The diameter of 2.0 mm is constant throughout it's length.

Remarks: The specimen may be distinguished from other ichnospecies of Spirophycus by it's lack of ornament and lower number of initial spirals (see Książkiewicz 1977). There are no associated ichnofossils.

Distribution. Aberystwyth Grits Formation

## 8.6 Palaeoecology

The concept of recurring and repeatable ichnofossil associations was first promoted by Seilacher (1967) who suggested that the distribution of these ichnofacies was a function of depth-related parameters. More recent recognition of the complexity of the situation (e.g. Ekdale and Mason 1988) does not, however, detract from the usefulness of the ichnofacies model in testing the environmental significance of the ichnofaunal assemblages. In the following sections the distribution of the ichnofaunal assemblages in the study area is examined, both in terms of the ichnofacies model of Seilacher (1967) and also in terms of their ethological groupings and the depositional environments in which they occur.

### 8.6.1 Ichnofaunal distribution patterns within the Llandeilo-Llandovery succession

Each of the ichnogenera from the southern Welsh Basin has a known (in so far as the literature allows) range of environments within which it may occur. Figure 8.1 shows the ichnofacies model of Seilacher (1967) with the probable ranges of the 15 ichnogenera found in the study area superimposed. From the diagram it can be seen that the ichnofacies, within which the majority of the ichnogenic ranges overlap, is that of the Nereites ichnofacies.

Two of the noted ichnogenera (Cosmorhaphe, Desmograpton) have only been recorded from documented probable deep-water environments. A further five ichnogenera (Circulichnis, Helminthoida, Paleodictyon, Protopaleodictyon, Spirophycus) are considered eurybathic although they are most common in deep-sea environments. Thus seven of the ichnogenera recorded in the area are rarely found in depositional environments which are not deep marine. However, just over half (8) of the ichnogenera noted in the southern Welsh Basin succession are eurybathic and are, therefore, not useful for ichnofacies determination. Given that almost half of the observed ichnogenera are rarely recorded outside of the Nereites ichnofacies, it may be concluded that the site of deposition for the study area was located in relatively deeper waters.

Four major sedimentary facies associations are recognised in the area and with respect to previously described analogues (for reviews, see Stow and Piper 1984; Pickering et al. 1986; Mutti and Normark 1987), each facies association is interpreted in terms of its depositional environment within a marine setting. The facies associations described herein are slightly different from those outlined in chapter four, with the lobe fringe facies being subdivided into two groupings (anoxic hemipelagites and mudstone turbidites) and the slope facies being grouped together with the sandstone lobe association as sheet system turbidites. This is in order to emphasise the possible controls, for example anoxia, on ichnofaunal distribution patterns.

While the detailed descriptions of these facies associations

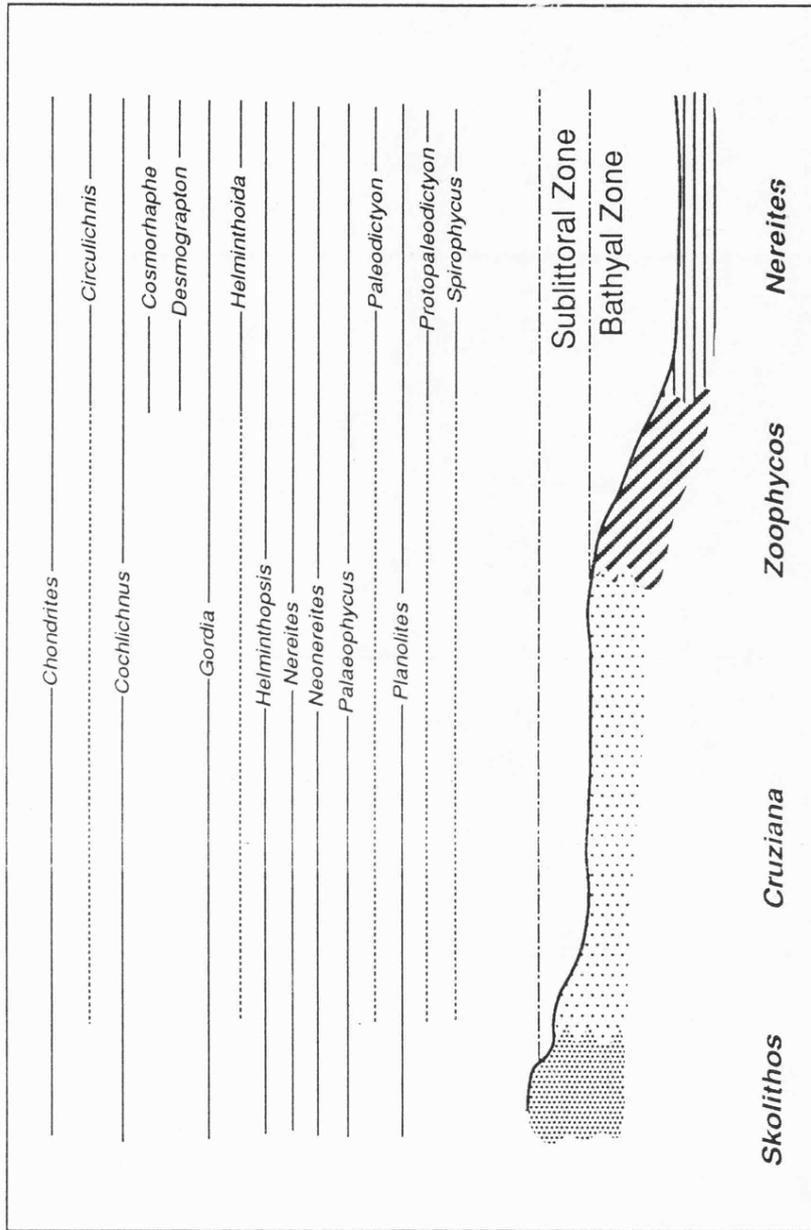


Figure 8.1 Schematic diagram showing the 15 collected ichnogenera and their ichnofacies ranges. (Solid lines indicate definite occurrence; dotted lines indicate possible occurrence.)

and their interpreted depositional environments are outside the scope of this chapter they can be summarised as follows (Fig 8.2).

(1). Anoxic distal shelf. - Hemipelagic, banded mudstones interbedded with laterally continuous and discontinuous silts and silty sands (Facies Class D). A single observation of possible bimodal current directions suggests that at least part of this environment was exposed to current reworking, albeit infrequently. This would suggest that the depositional environment was that of a distal shelf. The ichnofauna consists of a single specimen of the burrow of an infaunal deposit-feeding organism Planolites montanus. The presence of Planolites would suggest that levels of interstitial organic matter were high, a fact supported by the relatively high organic carbon values deduced from geochemical analysis of samples from the formation.

(2). Sheet system turbidites. - Turbiditic sandstones, siltstones and mudstones with minor conglomerates (Facies Classes A, B, C and D). These were formed as lobe-type deposits prograding from point sources to the south-west (McCann and Pickering 1989). The lateral continuity of the sandstones, together with the lack of evidence of major channel features, suggests that they were deposited as sheet systems (sensu Mutti and Normark 1987). Recognition of minor channels (possibly tributary channels) and fringe deposits on the margins of the major sandstone bodies confirms their lobe-like nature. These sheet-like turbidites contain the most abundant and diverse ichnofaunal assemblage noted in the succession. It should, however, be noted that the ichnofossils are not evenly distributed throughout the facies association, particularly in the slumped slope facies in part of the Llangranog Formation. This suggests a degree of heterogeneity within the lobe environments.

The presence of abundant horizontal burrows of deposit-feeding organisms (e.g. Cosmorhappe sinuosa, Desmograpton fuchsi, Gordia marina, Protopaleodictyon submontanum) suggest that energy levels, over much of this

Ichnospecies	Depositional environment			
	Anoxic distal shelf	Sheet system turbidites	Mudstone turbidites	Anoxic hemipelagites
<i>Chondrites</i> ichnospp.		●	●	●
<i>Circulichnis montanus</i>		◐		
<i>Cochlichnus anguineus</i>		○		
<i>Cosmorhapse sinuosa</i>		○		
<i>Desmograpton fuchsi</i>		○		
<i>Gordia marina</i>		●		
<i>Helminthoida</i> ichnosp. nov. A		○		
<i>Helminthopsis abeli</i>		○		
<i>H. hieroglyphica</i>		○		
<i>H. granulata</i>		○		
<i>H. cf. irregularis</i>		○		
<i>Helminthopsis</i> ichnosp.		●		
<i>Neonereites uniserialis</i>		○		
<i>Nereites cf. macleayi</i>		○		
<i>Palaeophycus anulatus</i>		○		
<i>P. tubularis</i>		●		
<i>Palaeophycus</i> ichnosp.		●		
<i>Paleodictyon (Gleno.) imperfectum</i>		●		
<i>P. (Squamo.) petaloideum</i>		●		
<i>Planolites montanus</i>	○	●	●	●
<i>Protopaleodictyon cf. incompositum</i>		◐		
<i>P. submontanum</i>		◐		
<i>Spirophycus bicornis</i>		○		

Figure 8.2 Environmental distribution and abundance data of the 23 ichnospecies recorded from the study area.

facies association, were relatively low. This is confirmed by the preservation of delicate markings on a number of ichnospecies, for example the transverse annulation on Palaeophycus anulatus which suggests that the sediment was stable and cohesive. The variety of ichnofossils preserved would suggest that levels of interstitial organic detritus were high and that oxygen levels were sufficient to support the relatively large population of trace-making organisms. Suspended organic matter was also relatively abundant as deduced from the presence of Palaeophycus tubularis, the burrows of filter-feeding organisms.

The least diverse and abundant ichnofaunal assemblages are found in areas where channel and channel margin facies predominate, for example the Poppit Sands Formation and the Aberystwyth Grits Formation in the vicinity of New Quay. In these areas it is likely that the erosive nature of the sandstone bodies would remove surface and shallow infaunal traces.

(3). Mudstone turbidites. - Siltstone and rare fine-grained sandstone turbidites interbedded with turbiditic mudstones (Facies Class D and E). The mudstone turbidites are frequently situated on the margins of major sandstone lobes and as such may be coeval with some of the sandstone bodies representing lobe fringe deposits. The extreme lateral continuity of some of the mudstones, together with their position vis-a-vis the sheet sandstones suggests that the mudstones are "distal" equivalents of some of the sandstones. Mudstone turbidites were deposited during periods of higher sea level stands and in these cases represent periods of reduced sediment input to the depositional system. Two ichnospecies are recorded from this environment, namely Chondrites ichnosp. and Planolites montanus. The presence of these infaunal deposit feeders would suggest that levels of interstitial organic matter were high.

(4). Anoxic hemipelagites. - These intercalated laminated hemipelagites and mudstone turbidites (Facies Class E) were deposited during periods of high eustatic sea-level stands.

The main deposit of this nature in the study area was deposited at the Ordovician-Silurian boundary, immediately following the late-Ashgill fall in sea-level. The laminated hemipelagites are rich in organic carbon and pyrite and exhibit a number of features consistent with deposition in anoxic, or low oxygen, environments. Two ichnospecies are recognised, namely Chondrites ichnosp. and Planolites montanus, albeit in greatly reduced numbers compared with the oxic mudstone turbidites. Their relative rarity would suggest that the depositional environment was not altogether conducive to the presence of deposit-feeding infaunal organisms.

It should, however, be noted that many of the ichnogenera (e.g. Cochlichnus, Cosmorhappe, Paleodictyon, Spirophycus) are not evenly distributed throughout the area and are instead restricted to the sheet system facies association (Fig. 8.2). Thus while this may be confidently assigned to a Nereites ichnofacies the other facies associations do not reflect a strong signature from any particular ichnofacies.

#### 8.6.2 Ethology of the ichnofaunal assemblage

Further information may be derived from examination of the behavioural, or ethological, groupings of the ichnofossil assemblage. Classification of ichnofossils according to the inferred behaviour of the producing organism was first proposed by Seilacher, in a series of publications (1953, 1964a, 1967b), and updated by Frey and Seilacher (1980). The ichnogenera of the southern Welsh Basin may be referred to one of four of these ethological groups. Fodinichnia represent traces of infaunal deposit feeders combining dwelling and sediment processing. Pascichnia are grazing trails which combine locomotion and feeding. Agrichnia are structures with horizontal tunnels arranged in regular geometric patterns whose ethologic significance is unclear (Ekdale et al. 1984). Domichnia are permanent dwelling burrows of infaunal organisms.

<b>Ichnogenus</b>	<b>Ethological group</b>	<b>Producer</b>
<i>Chondrites</i>	D, F	Annelid, Siphunculid, Polychaete, Sea pen, Arthropod
<i>Circulichnis</i>	?D, ?F	Polychaete, Annelid
<i>Cochlichnus</i>	P	Polychaete, Annelid, Nematode
<i>Cosmorhapha</i>	P	Polychaete, Acorn worm
<i>Desmograpton</i>	P, ?D	Polychaete, vermiform organism
<i>Gordia</i>	P	Polychaete, Gastropod, vermiform organism
<i>Helminthoida</i>	P	Polychaete, vermiform organism, Enteropneust
<i>Helminthopsis</i>	P	Polychaete, Gastropod, Priapulid annelid, vermiform organism
<i>Neonereites</i>	P	Vermiform organism
<i>Nereites</i>	P	Vermiform organism, Gastropod, Sea pen, Crustacean
<i>Palaeophycus</i>	D	Polychaete, Gastropod, Pelecypod, Annelid, vermiform organism
<i>Paleodictyon</i>	A	?Polychaete, ?vermiform organism
<i>Planolites</i>	F	Polychaete, Enteropneust, vermiform organism
<i>Protopaleodictyon</i>	P	Polychaetes, Annelids, vermiform organism
<i>Spirophycus</i>	P	Polychaete, Gastropod, Acorn worm

Figure 8.3 Ethological groupings and possible producing organisms for the 15 ichnogenera recorded from the study area.

The Nereites ichnofacies normally contains a high diversity of pascichnia and agrichnia. Most of the ichnogenera recorded from the succession belong to one of these two ethological groups (Fig. 8.3). Thus on the basis of the ethology of the ichnogenera the ichnofaunal assemblage of the study area can be assigned to the Nereites ichnofacies (Seilacher 1967).

### 8.7 Conclusions

In summary, the ichnofaunal assemblage from the southern Welsh Basin may be confidently assigned to the Nereites ichnofacies of Seilacher (1967). The range of ichnofossils recorded in the area include a number which have only ever been recorded from deep-marine depositional settings together with a greater number which are commonly considered to be typical of deep-marine settings. Furthermore, the dominance of pasichnial and agrichnial traces, within the assemblage would tend to confirm the assemblage as a typical Nereites one.

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## CHAPTER NINE - THE ICHNOGENUS DESMOGRAPTON FROM THE SILURIAN OF WALES - FIRST RECORD FROM THE PALAEOZOIC

### Abstract

The ichnogenus Desmograption is described from the Upper Llandovery age Aberystwyth Grits Formation in Wales. This is the first record of Desmograption both from the Palaeozoic and from Britain.

### 9.1 Introduction

The ichnogenus Desmograption Książkiewicz, 1977 is described for the first time from the Palaeozoic. The ichnogenus has previously been noted from strata of Mesozoic age in Europe and North and South America (e.g. D'Alessandro, 1980; Książkiewicz, 1977; Macsotay, 1967; McCann and Pickerill, 1988). This new occurrence is also significant because it extends the geographic range of Desmograption into Britain.

It has been generally accepted that deep-sea ichnofaunal assemblages of Palaeozoic age are less diverse than those of Mesozoic and younger strata (Crimes, 1977; Frey and Seilacher, 1980; Seilacher, 1974). Crimes (1973) noted the large and diverse numbers of ichnogenera present in Palaeozoic shallow marine environments (e.g. Crimes, 1970a; Banks, 1970) and interpreted the anomaly as resulting from the lack of colonisation of the deep sea environment in Palaeozoic times. Recent examination, however, of a selection of studies from the Palaeozoic has shown that the ichnoassemblages are not as restricted as was once thought (McCann and Pickerill, 1988). In fact many ichnogenera have quite a wide stratigraphic range, particularly those ichnogenera which have a generalized morphology and could thus be produced by a wide range of organisms.

## 9.2 Location and stratigraphy

The specimen of Desmograptus was found at Aberystwyth (SN 583827) within a sequence comprising interbedded sandstones, siltstones and mudstones in the upper part of the Silurian Aberystwyth Grits Formation. The Aberystwyth Grits Formation was first described by Keeping in 1878, with subsequent workers concentrating on the graded bedding which is well exhibited throughout (Rich, 1950; Kuenen, 1953). Smith (1956), and later Wood and Smith (1959), carried out an extensive study of the sedimentation and sedimentary history of the Aberystwyth Grits Formation and its lateral equivalents. They placed the Formation in the Monograptus turriculatus and M. crispus zones of the upper part of the Llandovery Series.

## 9.3 Sedimentary environment

The Aberystwyth Grits Formation is now interpreted as part of a relatively deep-water fan association. The recognition of canyon heads, large channels and coarse inner fan deposits around the trough-platform marginal zone is comparable with the situation as observed in Recent analogues (Bassett, 1984). According to Cave (1979) the basinal nature of the region is revealed in its sedimentary types, sedimentary structures and faunal content and by contrast with the contemporaneous shelf and reef deposits to the east in the Welsh Borderlands and Shropshire (c.f. Bassett, 1984; Cummins, 1969; Kelling and Woollands, 1969). Turbidity currents were suggested as the depositional mechanism for the sediments of the Aberystwyth Grits Formation with the material being brought in from a shallow-water source area at the Southern Basin margin (present day Pembrokeshire). It should, however, be noted that there is some evidence of a northern structural high during the Late Llandovery (McCann and Pickering, 1989). Preliminary petrographic analysis of coarse sandstones from the Aberystwyth Grits Formation, carried out by the author, notes the presence of felsic lithic fragments which are derived from the Ordovician

volcanic rock piles to the south in the Fishguard and Trefgarn areas.

The Formation contains a rich ichnofaunal assemblage which has not been described systematically but has featured in a number of accounts, notably Crimes and Crossley (1980) and Crimes (1970b). Associated ichnofossils, occurring on the same bedding surface as Desmograpton fuchsi, include Circulichnis montanus, Helminthopsis ichnosp., Palaeophycus tubularis, Palaeophycus ichnosp. and Paleodictyon (Glenodictyon) imperfectum. This note forms part of a larger systematic study of the Aberystwyth Grits Formation and related rocks to the southwest.

#### 9.4 Systematic ichnology

Ichnogenus Desmograpton Fuchs, 1895

1895 Desmograpton Fuchs, pp. 394-395, pl. 5, figs. 1-2, 4-6.

1967 Pseudodesmograpton Macsotay, p. 36

non 1971 Baroccoichnites Vialov, p. 88

1977 Desmograpton (Fuchs). Książkiewicz, p.181

non 1978 Desmograpton Radwanski, p. 56, plate 2, fig. 1-3.

Type ichnospecies: Desmograpton fuchsi Książkiewicz, 1977 by subsequent designation.

Diagnosis: Biramous burrows with lateral appendages along a straight or undulating axis in which either the transverse or longitudinal elements are pronounced (after Seilacher, 1977).

Discussion: In his original description of the ichnogenus Fuchs (1895) did not use any species level name to differentiate the ichnofossils he illustrated (pl. 5, figs., 1-2, 4-6). Subsequently one of the illustrations (pl. 5, fig. 2) was chosen as the type ichnospecies (Książkiewicz, 1977) and named Desmograpton fuchsi. The diagnosis is based on the material collected by Książkiewicz from the Polish Carpathians rather than on the original type material of

Fuchs.

Książkiewicz has erroneously been cited as the author of the ichnogenus (e.g. Książkiewicz, 1977). Fuchs, as the first person to describe the ichnogenus, is the correct author (e.g. Hantzschel, 1975) in accordance with ICZN rules.

The presence of transversal links is a major diagnostic feature of Desmograpton (Crimes et al., 1981). Conversely the absence of these links would result in an ichnofossil which could be a central portion of either Helminthoida or a Urohelminthoida. In such cases a median swelling of the burrows would help to make a positive identification (Książkiewicz, 1977). Desmograpton pieninicus Radwanski, 1978 is a case in point where the absence of both transversal links and median swelling would suggest that the form does not belong to the ichnogenus Desmograpton.

A number of authors (e.g. Hantzschel, 1975; Seilacher, 1977) have placed Pseudodesmograpton Macsotay, 1967 into synonymy with Desmograpton. Based on the original diagnosis of Macsotay (1967) Pseudodesmograpton comprises parallel burrows with occasional cross links. Macsotay also notes, however, that Pseudodesmograpton ichthyformis "...differs from Desmograpton in that it has a large central cylinder in place of small alternating burrows." (trans. litt. Macsotay, 1967, p.36). It would therefore appear that Macsotay (1967) was incorrect in erecting a new ichnogenus based on a minor change in the positioning of the transverse links and thus the synonymy holds. It would, however, be best to use the ichnospecific term Desmograpton fuchsi for specimens showing alternating transversal links and D. ichthyformis for specimens showing a large central cylinder. This is in keeping with the original figures of D. fuchsi (Fuchs, 1895).

Baroccoichnites Vialov, 1971 was placed into synonymy with Desmograpton by Seilacher (1977). The original diagnosis, however, notes that the ichnofossil "...comprises two rows of curved burrows with lateral appendages which are arranged in a chessboard fashion..." (trans. litt., Vialov, 1971, p. 88).

While this diagnosis is similar to that of Desmograpton, the illustration (pl. 1, figs. 2, 3) reveals that Baroccoichnites Vialov, 1971 is in fact a rudimentary meandering trace with only the apical bends preserved, as noted by Książkiewicz (1977). Specimens of Baroccoichnites have also been reported and figured by Yang (1986) which are morphologically very distinct from Desmograpton.

Desmograpton is characteristic of deep water turbiditic environments (D'Alessandro, 1980; Vogeltanz, 1971) and ranges in age from lower Silurian (this study) to Tertiary (Hantzschel, 1975). It is a grazing trail probably formed by polychaete worms (Książkiewicz, 1977).

Desmograpton fuchsi Książkiewicz, 1977

Pl. 9.1, Fig. a, b

- 1895 Desmograpton Fuchs, pp. 394-395, pl. 5, figs. 1-2, 4-6.
- 1954 Desmograpton Fuchs; Seilacher, p. 218, fig. 14.
- 1959 Desmograpton Fuchs; Seilacher, p. 1069, fig. 15.
- non 1967 Pseudodesmograpton ichthyiformis Macsotay, p. 36, pl. 6, fig. 20.
- ?1967 Helminthopsis Heer; Macsotay, pl. 6, fig. 20.
- 1977 Desmograpton ichthyforme (Macsoyay); Seilacher, pp. 311-312, fig. 7d.
- 1977 Desmograpton inversum Seilacher, p. 312, fig. 7f.
- 1977 Desmograpton fuchsi Książkiewicz, pp. 182-183, pl. 29, fig. 5, text fig. 43.

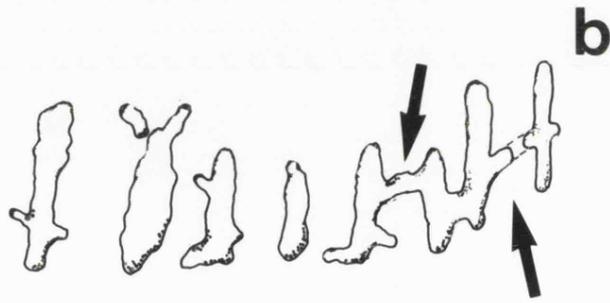
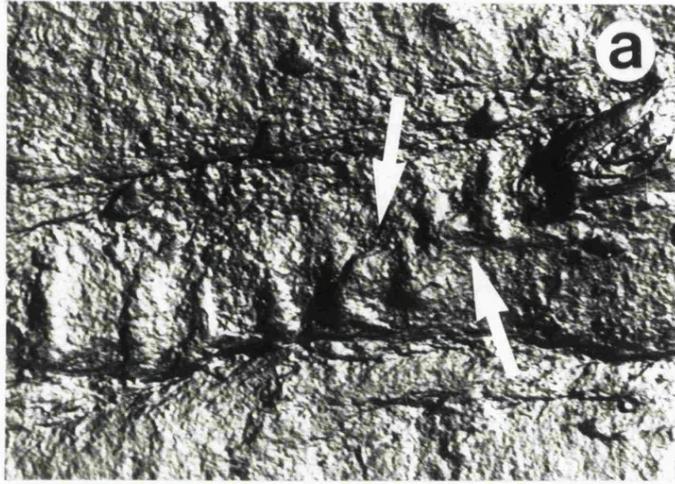
Diagnosis: Hypichnial, roughly straight and parallel burrows with some transversal links (Książkiewicz, 1977).

Material: One specimen (NMW89.12G.1) held in the National Museum of Wales, Cardiff.

Description: Smooth burrow system preserved in convex hyporelief on the sole of a 6.3cm thick, Tcde (Bouma, 1962),

Plate 9.1

- a. Desmograpton fuchsi preserved on a lower bedding surface in convex hyporelief, Aberystwyth Grits Formation (Upper Llandovery), Aberystwyth (SN), NMW89.12G.1, x 1.85.
- b. Outline drawing of the same specimen. The diagnostic transversal links are arrowed.



fine- to medium-grained sandstone. The overall length of the structure is 48.0mm, with burrow diameter ranging between 0.3-1.7mm. Some of the parallel to sub-parallel burrow elements show pronounced medial thickening. The transverse links are arranged alternately and range from 2.0-5.0mm in length and from 0.5-2.3mm in diameter.

Remarks: The specimen contains more than one transversal link and hence is confidently referred to the ichnospecies Desmograpton fuchsi which may be distinguished from D. ichthyformis by the presence of alternating transversal links.

In his discussion of the ichnogenus Seilacher (1977) erected three ichnospecies Desmograpton ichthyforme (= Pseudodesmograpton ichthyformis Macsotay, 1967), D. geometricum and D. inversum. Of these, only Desmograpton geometricum is herein deemed to be a viable ichnospecies. Desmograpton ichthyforme is not a suitable choice of name since it is an objective synonym of the ichnospecies described by Macsotay (1967) rather than with the original figures of Fuchs (1895). As noted in the discussion of the ichnogenus, the ichnospecific term Desmograpton ichthyformis Macsotay, 1967 is more properly used to refer to ichnospecies having a large central cylinder. Desmograpton inversum has been placed into synonymy with D. fuchsi as it is felt that the criteria used by Seilacher (1977) to distinguish it from related ichnospecies are too vague to be of any great systematic value. It should also be noted that Książkiewicz (1977, fig. 43g) figures a specimen of Desmograpton fuchsi which is identical to the illustration of D. inversum as figured by Seilacher (1977, fig. 7f).

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**CHAPTER TEN - CONCLUSIONS****10.1 Conclusions**

1. The Llandeilo-Llandovery succession between Dinas Head and Cardigan was mapped and four formations (Parrog, Newport Sands, Ceibwr and Poppit Sands) were established on the basis of lithostratigraphy.
2. These four formations, together with the overlying ten formations (previously identified by Hendricks 1926; Smith 1956; Anketell 1963; Craig 1985) are described in detail and representative logged sections are presented.
3. The sediments of the succession are interpreted as representing distal shelf, sandstone lobes, lobe fringes and slope environments; sub-environments of a major turbidite system (or systems). The turbidite systems developed in the Welsh depositional basin as a Type 1 (unchanneled sandstone lobes) system. Lobe development is closely linked to global eustatic sea-level variations with the main periods of sandstone lobe progradation coinciding with lower sea levels. Tectonic influences, however, are locally important.
4. The Sea of Okhotsk (Western Pacific Ocean) is a useful modern analogue for the Lower Palaeozoic Welsh Basin. Its geomorphology comprising a shelf, slope and trench situated behind an oceanic island arc is similar, in many respects, to that of the Lower Paleozoic Welsh Basin which was situated to the south of the Leinster-Lake District volcanic arc.
5. The provenance of Llandeilo to upper Llandovery sediments from the Welsh Basin has been investigated by petrographic and geochemical methods. Sandstone-dominated formations were deposited from high-concentration turbidity currents. Petrographic data suggest derivation from a non-collisional recycled or cratonic setting. This is broadly confirmed by analysis of microconglomerate lithic fragments and the major

element geochemistry. Mudstone-dominated formations were deposited from dilute turbidity currents and as hemipelagic sediments. The mudstone geochemistry indicates a tectonic setting transitional between an active continental margin and a passive margin. This study demonstrates that neither set of analytical methods are individually adequate for provenance reconstruction and it is advisable to use a variety of techniques for greater confidence in interpretation.

6. Laminated hemipelagites are recognised from Llandeilo-Llandovery age sediments within the Welsh Basin. X-ray diffraction and whole rock geochemical analyses failed to distinguish the hemipelagic sediments from associated mudstones and siltstone turbidites. Internally the hemipelagites are characterised by laminae of pyrite framboids. These laminae do not show any evidence of primary deposition and are interpreted as diagenetic products formed under euxinic or semi-euxinic conditions.

7. The strata contain a diverse and relatively abundant ichnofaunal assemblage consisting of fifteen ichnogenera: Chondrites, Circulichnis, Cochlichnus, Cosmorhappe, Desmograpton, Gordia, Helminthoida, Helminthopsis, Nereites, Neonereites, Palaeophycus, Paleodictyon (Glenodictyon), Paleodicyton (Squamodictyon), Planolites, Protopaleodictyon and Spirophycus. The ichnogenera are unevenly distributed throughout the succession, the main controlling factors on their distribution being toponomy, anoxia, and a global extinction event. The relative importance of these controls is examined with reference to 33 previously described and taxonomically well-documented deep-water flysch ichnofossil assemblages covering the Cambrian to Tertiary worldwide.

8. Twenty three ichnospecies are described, two of which (Desmograpton fuchsi and Spirophycus bicornis) are noted for the first time from the Palaeozoic. Sheet sands contain the most abundant and diverse ichnoassemblages, a reflection of favorable environmental conditions for inhabitation by benthic organisms. Oxidic turbiditic mudstones contain just

two ichnospecies, namely Chondrites ichnosp. and Planolites montanus both of which are common. Anoxic hemipelagites contain the same two ichnospecies albeit in greatly reduced numbers, while the hemipelagites deposited in a distal shelf environment contain the least abundant and diverse ichnoassemblage, comprising just a single specimen of Planolites montanus.

## 10.2 Suggestions for further work

1. The Ordovician and Silurian stratigraphies to the north of the Aberystwyth Grits Formation could be integrated with those which outcrop between Dinas Head and Aberystwyth. This might provide a more complete picture of depositional events within the Welsh Basin during these times.
2. The petrographic data for the study area gave a very uniform reading for the entire period of deposition, suggesting derivation of much of the material from the Precambrian basement. The tectonic setting thus suggested was, therefore a relict one, rather than being an accurate reflection of what was actually happening in the Welsh Basin throughout the period of deposition. Whether this is true for the Welsh Basin as a whole has not yet been determined.
3. The petrographic database for the Welsh Basin is poor overall. What is required is to examine a large number of samples over a wide area, both stratigraphically and areally to try and constrain the source area. Problems associated with sediment reworking could then, hopefully, be eliminated.
4. Similarly, the geochemical database for sediments is very poor. This must be enlarged if any meaningful geochemical models are to be compiled. The major problem with sedimentary geochemistry would appear to be the extreme variability of many elements in depositional and diagenetic systems. Thus any models would have to account for these variations. Certainly the models of Bhatia (1983, 1985) are extremely unreliable, although they could possibly be

improved with greatly increased databases and more judicious sampling.

5. Laminated hemipelagites have been recognised by a number of workers but still remain poorly understood. Much more work, both descriptive and analytical, is required before they can be definitively interpreted. The methods used by Dimberline (1987), while interesting, have very limited applicability. Perhaps the way forward is merely to describe and catalogue sufficient numbers of them (from both the Recent and the ancient), in order to be aware of the degree of variation, before trying to produce models to account for their formation.

6. While a large number of detailed ichnofossil taxonomic studies have been identified worldwide, there are a significant and important number of gaps. These include the relative lack of work on Permian and Triassic deep-water ichnofossil assemblages, the preponderance of work in North America and Europe and the corresponding lack of work in Africa and South America. These gaps can only be rectified by systematic fieldwork, collecting and description.

7. In recent years ichnofossil assemblages have been examined more rigorously than before. This is particularly true of the work that has been done linking different ichnofossils with levels of oxygen in the bottom waters (e.g. Savrda & Bottjer (1987)). It is important that taxonomic studies do not ignore the palaeoecologic dimension given that it is this which elevates the significance of ichnofossil occurrences.

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APPENDIX ONE - GEOCHEMICAL DATA FROM THE WELSH BASIN

Sandstone Geochemistry

Pellet N	L11008	L11584	L11594	L11605	L10986	L11600	L11582	L10980
Sample N	T121	T120X	T79	T86	T171	TW2	TW3	T45
Location	NS	PS						
SiO2	70.5	78.0	94.4	87.1	75.8	83.3	79.4	69.7
TiO2	0.62	0.34	0.27	0.37	0.40	0.36	0.34	0.66
Al2O3	15.6	12.2	6.9	10.5	12.4	11.5	10.8	14.8
Fe2O3	0.6	0.8	0.4	0.6	0.5	0.6	0.6	0.7
FeO	ND							
MnO	0.04	0.04	0.01	0.05	0.02	0.03	0.04	0.04
MgO	1.8	1.9	0.1	2.0	0.8	1.6	2.5	1.9
CaO	0.3	0.1	0.1	0.4	0.2	0.1	0.2	0.2
Na2O	1.3	0.3	0.4	0.6	1.4	1.5	0.5	1.8
K2O	2.59	0.39	2.26	1.15	2.26	1.68	2.89	1.82
P2O5	0.38	0.13	0.05	0.37	0.24	0.14	0.25	0.18
LOI	ND							
Total	93.78	94.25	105.04	103.26	94.05	100.81	97.68	91.86

Trace elements in parts per million

V	98	63	31	56	63	52	48	88
Cr	59	59	19	34	37	24	56	125
Ba	406	453	280	528	446	336	354	327
La	21	BDL	BDL	BDL	21	BDL	BDL	16
Ce	47	BDL	BDL	BDL	52	BDL	BDL	31
Nd	22	BDL	BDL	BDL	31	BDL	BDL	14
Nb	8	6	5	ND	5	ND	ND	10
Zr	140	83	77	ND	92	ND	ND	186
Y	28	36	17	ND	23	ND	ND	25
Sr	63	57	36	ND	71	ND	ND	64
Rb	72	12	40	ND	52	ND	ND	66
Th	10	8	5	ND	7	ND	ND	8
Ga	14	8	7	ND	17	ND	ND	14
Zn	47	51	5	ND	78	ND	ND	63
Ni	16	14	3	ND	18	ND	ND	20
La/Th	2.0	—	—	—	2.8	—	—	2.0
Rb/Sr	1.2	0.2	1.1	—	0.7	—	—	1.0
Nb/Y	0.3	0.2	0.3	—	0.2	—	—	0.4
Zr/Nb	16.9	14.9	16.7	—	18.7	—	—	19.3

## Appendix.2

## Sandstone Geochemistry

Pellet N	L10981	L11009	L11010	L11599	L11602	L10984	L11583	L11586
Sample N	T11	T45	T62	T91	T92	T103	T114	T115
Location	PS							
SiO <sub>2</sub>	80.7	73.0	64.2	81.7	88.8	70.6	72.7	72.8
TiO <sub>2</sub>	0.39	0.51	0.84	0.31	0.30	0.49	0.54	0.50
Al <sub>2</sub> O <sub>3</sub>	10.6	13.0	21.7	10.5	8.5	12.7	15.7	12.6
Fe <sub>2</sub> O <sub>3</sub>	0.7	0.7	0.5	0.5	0.6	1.1	0.6	1.3
FeO	ND							
MnO	0.05	0.04	0.02	0.04	0.04	0.13	0.02	0.05
MgO	2.0	2.2	1.4	1.5	1.1	2.4	1.1	3.5
CaO	0.2	0.3	0.0	0.4	0.3	0.3	0.6	0.3
Na <sub>2</sub> O	1.4	1.6	1.4	1.3	1.7	1.8	1.4	0.6
K <sub>2</sub> O	1.20	1.35	3.58	1.88	0.68	1.53	2.14	1.31
P <sub>2</sub> O <sub>5</sub>	0.14	0.23	0.18	0.34	0.31	0.30	0.58	0.47
LOI	ND							
Total	97.39	92.92	93.91	98.51	102.39	91.38	95.34	93.37

## Trace elements in parts per million

V	64	66	134	51	40	75	91	88
Cr	48	73	85	56	33	47	73	44
Ba	208	225	864	184	140	148	728	412
La	17	15	19	BDL	BDL	16	BDL	BDL
Ce	38	32	38	BDL	BDL	35	BDL	BDL
Nd	17	16	19	BDL	BDL	17	BDL	BDL
Nb	7	8	13	ND	ND	9	8	7
Zr	92	145	190	ND	ND	160	138	125
Y	23	23	28	ND	ND	27	34	31
Sr	55	51	103	ND	ND	49	90	220
Rb	39	46	118	ND	ND	44	69	44
Th	5	8	10	ND	ND	8	3	8
Ga	10	13	20	ND	ND	11	12	14
Zn	61	74	46	ND	ND	101	42	105
Ni	15	19	20	ND	ND	39	19	30
La/Th	3.2	1.8	2.0	—	—	2.0	—	—
Rb/Sr	0.7	0.9	1.1	—	—	0.9	0.8	0.2
Nb/Y	0.3	0.4	0.5	—	—	0.3	0.2	0.2
Zr/Nb	13.8	18.4	14.6	—	—	18.8	17.3	16.9

## Appendix.3

## Sandstone Geochemistry

Pellet N	L11585	L11593	L11596	L11581	L11604	L10978	L10978	L11580
Sample N	T68	T67	T69	T70	T66	TT2	TT2	T162
Location	LLAN	LLAN	LLAN	AGUCH	AGUCH	AG	AG	AG
SiO <sub>2</sub>	80.1	86.7	90.2	62.4	88.3	62.8	ND	64.7
TiO <sub>2</sub>	0.38	0.33	0.38	0.91	0.34	0.81	ND	0.95
Al <sub>2</sub> O <sub>3</sub>	11.3	9.4	9.2	20.4	9.2	17.8	ND	20.9
Fe <sub>2</sub> O <sub>3</sub>	0.7	0.6	0.6	1.2	0.6	1.3	ND	1.0
FeO	ND							
MnO	0.04	0.10	0.07	0.08	0.06	0.17	ND	0.05
MgO	1.6	1.1	1.1	2.7	0.9	2.3	ND	2.1
CaO	0.2	0.2	0.1	0.4	0.1	0.2	ND	0.1
Na <sub>2</sub> O	2.0	2.3	1.7	1.3	2.4	1.4	ND	1.3
K <sub>2</sub> O	1.01	0.70	0.85	3.18	0.62	2.67	ND	3.43
P <sub>2</sub> O <sub>5</sub>	0.14	0.10	0.11	0.28	0.09	0.14	ND	0.12
LOI	ND							
Total	97.38	101.57	104.41	92.73	102.64	89.49	0.00	94.59

## Trace elements in parts per million

V	46	39	39	130	35	114	ND	159
Cr	27	18	37	85	30	74	ND	92
Ba	142	104	133	434	88	384	ND	763
La	BDL	BDL	BDL	BDL	BDL	29	ND	BDL
Ce	BDL	BDL	BDL	BDL	BDL	66	ND	BDL
Nd	BDL	BDL	BDL	BDL	BDL	28	ND	BDL
Nb	7	5	7	15	ND	14	ND	16
Zr	124	110	139	219	ND	214	ND	225
Y	14	16	25	40	ND	31	ND	33
Sr	40	37	35	87	ND	73	ND	109
Rb	41	29	38	132	ND	107	ND	153
Th	7	4	7	14	ND	8	ND	14
Ga	9	10	9	24	ND	20	ND	25
Zn	56	52	49	114	ND	132	ND	102
Ni	17	11	18	39	ND	44	ND	37
La/Th	—	—	—	—	—	3.7	—	—
Rb/Sr	1.0	0.8	1.1	1.5	—	1.5	—	1.4
Nb/Y	0.5	0.3	0.3	0.4	—	0.4	—	0.5
Zr/Nb	17.5	22.4	21.4	14.5	—	15.8	—	14.2

## Appendix.4

## Sandstone Geochemistry

Pellet N	L11592	L11603	L11606	L11598	L10973	L11588	L11595	L11597
Sample N	T161	T159	T152	T167	T184	T194	T188	T190
Location	AG							
SiO <sub>2</sub>	68.3	74.4	71.0	87.9	65.1	85.3	65.1	62.8
TiO <sub>2</sub>	0.61	0.57	0.52	0.33	0.84	0.74	0.84	0.79
Al <sub>2</sub> O <sub>3</sub>	16.0	14.4	15.3	8.8	17.0	11.9	18.3	17.6
Fe <sub>2</sub> O <sub>3</sub>	0.9	0.7	0.8	0.6	0.9	0.5	1.2	1.1
FeO	ND							
MnO	0.21	0.06	0.12	0.06	0.03	0.11	0.03	0.11
MgO	2.3	2.0	2.7	1.0	2.1	1.1	2.6	2.5
CaO	1.0	0.9	1.5	0.2	0.2	0.1	0.2	1.7
Na <sub>2</sub> O	1.9	2.3	2.2	2.4	1.8	1.1	1.5	1.3
K <sub>2</sub> O	2.24	1.76	1.87	0.62	2.62	1.68	2.25	2.37
P <sub>2</sub> O <sub>5</sub>	0.11	0.11	0.13	0.10	0.15	0.09	0.15	0.14
LOI	ND							
Total	93.62	97.20	96.10	101.94	90.82	102.63	92.27	90.53

## Trace elements in parts per million

V	75	70	68	36	102	69	105	102
Cr	42	45	39	17	84	69	68	67
Ba	370	271	275	96	331	298	281	361
La	BDL	BDL	BDL	BDL	29	BDL	BDL	BDL
Ce	BDL	BDL	BDL	BDL	52	BDL	BDL	BDL
Nd	BDL	BDL	BDL	BDL	26	BDL	BDL	BDL
Nb	11	ND	ND	ND	15	12	15	13
Zr	195	ND	ND	ND	275	517	296	242
Y	25	ND	ND	ND	30	23	32	30
Sr	55	ND	ND	ND	62	45	60	66
Rb	79	ND	ND	ND	97	72	88	89
Th	9	ND	ND	ND	12	8	11	12
Ga	16	ND	ND	ND	18	13	20	20
Zn	71	ND	ND	ND	78	63	85	110
Ni	29	ND	ND	ND	35	20	47	38
La/Th	—	—	—	—	2.5	—	—	—
Rb/Sr	1.4	—	—	—	1.6	1.6	1.5	1.4
Nb/Y	0.4	—	—	—	0.5	0.5	0.5	0.4
Zr/Nb	18.6	—	—	—	18.7	43.8	20.3	18.6

## Appendix.5

### Sandstone Geochemistry

Pellet N L11587  
Sample N T89  
Location AG

SiO <sub>2</sub>	70.3
TiO <sub>2</sub>	0.73
Al <sub>2</sub> O <sub>3</sub>	16.0
Fe <sub>2</sub> O <sub>3</sub>	1.0
FeO	ND
MnO	0.15
MgO	2.4
CaO	1.2
Na <sub>2</sub> O	1.5
K <sub>2</sub> O	2.11
P <sub>2</sub> O <sub>5</sub>	0.16
LOI	ND
Total	95.48

### Trace elements in parts per million

V	91
Cr	68
Ba	269
La	BDL
Ce	BDL
Nd	BDL
Nb	12
Zr	205
Y	28
Sr	70
Rb	92
Th	10
Ga	20
Zn	83
Ni	36

La/Th	—
Rb/Sr	1.3
Nb/Y	0.4
Zr/Nb	16.8

Appendix.6

mudstone geochemistry

Pellet Sample Location	L10982 PA	L11001 T63 PA	L11002 T96 PA	L11005 T55 PA	L11007 T226 PA	L10974 T221 C	L10983 T225 C	L10992 T222 C
SiO2	64.4	53.5	58.1	68.7	58.5	63.3	68.0	64.5
TiO2	0.98	0.88	1.09	0.81	1.16	1.05	1.14	1.13
Al2O3	20.5	15.9	22.0	16.0	21.9	21.3	21.2	22.1
Fe2O3	7.0	20.3	7.7	10.4	9.2	8.4	2.7	4.7
FeO	ND	ND	ND	ND	ND	ND	ND	ND
MnO	0.06	0.22	0.07	0.08	0.07	0.07	0.02	0.02
MgO	1.9	5.4	2.5	2.0	1.9	2.0	1.2	1.4
CaO	0.1	0.5	0.2	0.6	0.1	0.1	0.0	0.1
Na2O	0.6	0.6	3.1	1.1	1.2	1.3	1.2	1.3
K2O	5.04	1.32	4.38	2.51	4.11	3.48	3.56	3.66
P2O5	0.09	0.42	0.12	0.53	0.23	0.10	0.12	0.04
LOI	ND	ND	ND	ND	ND	ND	ND	ND
Total	100.66	98.89	99.35	102.77	98.36	101.18	99.16	98.92

Trace elements in parts per million

V	163	153	170	127	193	163	186	191
Cr	99	96	115	79	129	99	134	117
Ba	494	264	575	343	684	487	2667	1288
La	29	32	36	43	32	24	34	28
Ce	61	71	68	99	66	45	73	54
Nd	25	29	29	49	28	19	31	24
Nb	15	14	19	15	19	18	16	18
Zr	200	203	237	187	239	223	212	195
Y	30	34	36	72	38	35	34	29
Sr	93	45	117	83	120	105	157	200
Rb	156	46	166	103	177	148	150	163
Th	11	13	16	11	13	15	13	14
Ga	30	25	28	21	28	26	29	22
Zn	94	245	113	122	95	108	25	35
Ni	35	64	42	45	36	39	5	8
La/Th	2.6	2.6	2.2	3.8	2.5	1.6	2.6	2.0
Rb/Sr	1.7	1.0	1.4	1.2	1.5	1.4	1.0	0.8
Nb/Y	0.5	0.4	0.5	0.2	0.5	0.5	0.5	0.6
Zr/Nb	13.0	14.7	12.8	12.7	12.6	12.5	13.0	10.9

## Appendix.7

## mudstone geochemistry from welsh basin

Pellet Sample Location	L10991 T41 PS	L10993 T215 PS	L10994 T54 PS	L10985 GW GWBERT	L10987 GW GWBERT	L10989 T214 GWBERT	L10996 T212 MWNT	L10979 T205 TRES
SiO <sub>2</sub>	63.8	61.2	61.1	64.7	60.9	61.9	69.0	61.0
TiO <sub>2</sub>	1.05	1.04	1.06	1.06	1.04	1.09	0.85	1.03
Al <sub>2</sub> O <sub>3</sub>	21.9	21.4	21.6	21.5	21.3	20.8	16.2	23.6
Fe <sub>2</sub> O <sub>3</sub>	7.3	8.8	8.2	6.1	8.6	8.3	7.3	7.2
FeO	ND	ND	ND	ND	ND	ND	ND	ND
MnO	0.08	0.13	0.08	0.07	0.09	0.05	0.08	0.03
MgO	1.9	2.1	2.1	1.7	2.3	2.2	1.8	1.8
CaO	0.1	0.1	0.2	0.2	0.3	0.2	0.2	0.1
Na <sub>2</sub> O	1.4	1.3	1.4	1.5	1.4	1.3	1.5	1.6
K <sub>2</sub> O	3.67	3.65	3.70	3.88	3.60	3.40	2.49	3.78
P <sub>2</sub> O <sub>5</sub>	0.07	0.11	0.21	0.12	0.24	0.11	0.11	0.07
LOI	ND	ND	ND	ND	ND	ND	ND	ND
Total	101.36	99.92	99.68	100.83	99.72	99.27	99.44	100.18

## Trace elements in parts per million

V	168	170	186	191	183	142	102	162
Cr	102	109	103	142	135	171	67	116
Ba	518	519	545	559	502	447	319	658
La	27	26	30	32	39	32	24	42
Ce	60	53	64	58	86	67	51	82
Nd	24	21	27	22	38	27	24	33
Nb	16	16	17	17	16	18	14	16
Zr	211	203	206	182	198	279	352	168
Y	32	33	35	37	46	31	27	31
Sr	107	92	108	109	101	74	58	125
Rb	156	153	151	167	151	131	91	187
Th	14	14	11	14	13	14	11	14
Ga	23	26	28	21	25	22	23	28
Zn	108	125	111	70	99	98	90	112
Ni	43	44	45	30	43	39	32	74
La/Th	1.9	1.9	2.7	2.3	3.0	2.3	2.3	2.9
Rb/Sr	1.5	1.7	1.4	1.5	1.5	1.8	1.6	1.5
Nb/Y	0.5	0.5	0.5	0.5	0.3	0.6	0.5	0.5
Zr/Nb	13.1	13.0	12.2	10.8	12.6	15.7	25.7	10.2

## Appendix.8

## mudstone geochemistry from welsh basin

Pellet Sample Location	L10988 T29 TRES	L10995 T208 TRES	L10998 T206 TRES	L10999 T210 TRES	L11003 T209 TRES	L11006 T204 TRES	L10990 T71 LLAN	L10997 T202 GAER
SiO <sub>2</sub>	62.9	62.0	60.8	61.2	63.1	60.9	61.5	61.2
TiO <sub>2</sub>	1.06	1.11	1.03	1.09	1.08	1.03	1.00	1.20
Al <sub>2</sub> O <sub>3</sub>	21.9	23.4	23.7	22.8	22.5	22.6	19.9	22.5
Fe <sub>2</sub> O <sub>3</sub>	9.1	7.3	7.7	8.5	8.8	6.8	9.1	9.5
FeO	ND	ND	ND	ND	ND	ND	ND	ND
MnO	0.09	0.06	0.04	0.08	0.07	0.02	0.07	0.06
MgO	2.0	1.7	2.1	1.9	1.7	1.7	2.7	1.8
CaO	0.1	0.2	0.3	0.1	0.1	0.2	0.2	0.2
Na <sub>2</sub> O	1.3	1.6	1.6	1.3	1.4	1.7	1.2	1.0
K <sub>2</sub> O	3.75	4.09	3.78	3.92	3.93	3.64	3.40	3.93
P <sub>2</sub> O <sub>5</sub>	0.06	0.07	0.07	0.07	0.07	0.08	0.10	0.24
LOI	ND	ND	ND	ND	ND	ND	ND	ND
Total	102.14	101.61	101.12	101.08	102.79	98.64	99.13	101.63

## Trace elements in parts per million

V	165	178	154	175	170	156	138	164
Cr	114	122	118	136	114	117	94	122
Ba	488	558	635	554	538	643	2286	655
La	38	41	39	40	44	44	31	47
Ce	86	80	74	81	87	95	62	100
Nd	37	29	28	29	36	51	24	43
Nb	18	19	17	18	18	17	16	23
Zr	178	183	170	174	171	174	185	217
Y	29	30	33	29	29	36	29	39
Sr	101	119	127	114	111	128	90	116
Rb	166	192	181	182	184	182	131	158
Th	15	12	13	15	16	14	10	17
Ga	26	32	30	30	29	27	24	37
Zn	118	105	110	119	115	98	122	126
Ni	45	41	45	45	49	92	43	44
La/Th	2.6	3.3	3.1	2.8	2.8	3.1	3.0	2.7
Rb/Sr	1.6	1.6	1.4	1.6	1.7	1.4	1.5	1.4
Nb/Y	0.6	0.6	0.5	0.6	0.6	0.5	0.5	0.6
Zr/Nb	10.0	9.8	10.0	9.6	9.7	10.0	11.9	9.5

## Appendix.9

## mudstone geochemistry from welsh basin

Pellet Sample Location	L11000 T200 LG	L11004 T201 CEFN	L10976 T170 GROGAL	L10977 T170 GROGAL	L10975 T184 AG
SiO <sub>2</sub>	60.8	62.0	63.6	60.7	61.6
TiO <sub>2</sub>	0.95	0.98	1.13	1.16	1.02
Al <sub>2</sub> O <sub>3</sub>	17.9	20.6	21.8	21.0	22.0
Fe <sub>2</sub> O <sub>3</sub>	10.7	9.0	7.6	8.3	7.1
FeO	ND	ND	ND	ND	ND
MnO	0.20	0.18	0.10	0.15	0.03
MgO	2.7	2.4	1.6	1.8	2.0
CaO	0.1	0.2	0.2	1.1	0.2
Na <sub>2</sub> O	1.6	1.5	1.3	1.4	1.2
K <sub>2</sub> O	2.47	3.40	4.08	3.76	4.39
P <sub>2</sub> O <sub>5</sub>	0.18	0.17	0.05	0.53	0.10
LOI	ND	ND	ND	ND	ND
Total	97.62	100.31	101.42	99.86	99.76

## Trace elements in parts per million

V	113	135	170	156	164
Cr	77	82	106	112	103
Ba	291	437	624	560	562
La	23	36	43	41	47
Ce	49	78	84	93	103
Nd	21	33	32	46	44
Nb	16	17	19	21	18
Zr	242	182	185	246	182
Y	24	33	30	47	36
Sr	74	89	112	116	75
Rb	92	137	169	146	181
Th	10	14	12	15	13
Ga	21	24	34	26	28
Zn	127	115	95	99	89
Ni	48	37	36	42	34
La/Th	2.3	2.7	3.5	2.8	3.6
Rb/Sr	1.2	1.5	1.5	1.3	2.4
Nb/Y	0.7	0.5	0.6	0.4	0.5
Zr/Nb	15.2	10.7	9.7	12.0	10.3

KEY: LOI - Loss on ignition; BDL - Below detection limit; ND - Not determined; AG - Aberystwyth Grits Formation; AGOCH - Allt Goch Formation; C - Ceibwr Formation; CEFN - Cefn Cwrt Formation; GAER - Gaerlwyd Formation; GWBERT - Gwbert Formation; LG - Llangraig Formation; LLAN - Llangranog Formation; MWNT - Mwnt Formation; NS - Newport Sands Formation; PA - Parrog Formation; PS - Poppit Sands Formation; TRES - Tresaith Formation.