SEDIMENTOLOGY OF THE LOWER AND MIDDLE CAMBRIAN OF NORTH WALES.

"Thesis submitted in accordance with the requirements of the University of Liverpool for the degree of Doctor in Philosophy by Christopher Johann Griffiths."

October, 1987.

CONTENTS.

SEDIMENTOLOGY OF THE LOWER AND MIDDLE CAMBRIAN OF NORTH WALES.

by C.J. Griffiths.

ABSTRACT.

The aim of this thesis is to examine in detail the sedimentary processes which were active during the deposition of the Lower and Middle Cambrian rocks of North Wales. Sediments are described from three main outcrop areas in the Harlech Dome, St Tudwal's Peninsula and Arfon. Sections are compared vertically within the succession and laterally between areas.

In the Harlech Dome the Cambrian succession is underlain by intermediate and basic volcanics. Above the volcanics there is a deepening trend up from fluvial to shallow marine and turbidite deposits. The turbidite deposits show a large scale alternation from sand-rich turbidite systems (Rhinog and Barmouth Formations) dominated by thick bedded, high density turbidity currents to silt/mud-rich turbidite systems (Llanbedr, Hafotty and Gamlan Formations) dominated by thin bedded sequences deposited from dilute turbidity currents. The Sand-rich turbidite systems were derived from the north (Rhinog Formation) and from the south (Barmouth Formation). They indicate deposition on weakly channelised lobes where channels have a very high width to depth ratios, are shallowly erosive, tend to be part of single scour-and-fill events, are probably distributed in a braided pattern and locally have convex upward tops (Compensation Cycles). Cross-bedding also occurs within these deposits indicating the presence of sustained low density, traction-dominated flows which reworked and transported volcanic-rich sediment derived from the east.

In St Tudwal's Peninsula the succession is similar except that the Cilan Grits (the lateral equivalent of the Barmouth Formation) were derived from the northeast. In general thick beds within the coarse turbidite formations are more laterally continuous than in the Harlech Dome and are indicative of turbidite lobe deposits.

In Arfon a thick sequence of subaerial, acidic ash-flow tuffs are overlain by laterally variable braided fluvial deposits, volcaniclastics and shallow marine deposits (Arfon Group) passing up into basinal turbidites. The basinal deposits are dominated by slates in their lower part (Llanberis Slates Formation) and indicate a different style of turbidite deposition to their probable lateral equivalent in the Harlech Dome: the sand-rich Rhinog Formation. The slates are overlain by the Bronllwyd Grit Formation which was deposited from mainly sand-rich, high density turbidity currents.

This study indicates that in general Lower and Middle Cambrian successions in North Wales show deepening trends which conform to the McKenzie (1978) model of lithospheric stretching (early rifting followed by more regional sagging). However there are considerable lateral variations in facies between the three main areas of outcrop. These differences suggest that for at least part of the Lower and Middle Cambrian sediments in the Harlech Dome- St Tudwal's Peninsula and Arfon were deposited in separate basins- the Merioneth and Arfon Basins respectively.

ACKNOWLEDGEMENTS.

I would like to thank the following people who assisted me with the research and writing of the thesis:

Drs. Pat Brenchley and Pete Crimes, my supervisors.

Steve Crowley for doing the isotope analyses and for his assistance in general.

Julie Sharman and Richard Oglethorpe for their encouragement.

Dave Oates and Pete Challis for preparing my thin sections.

Tim Hopkins and Dave Plant for the use of the electron microprobe at Manchester University.

I gratefully acknowledge the receipt of a N.E.R.C. studentship.

LIST OF FIGURES.

LIST OF PLATES.

CHAPTER ONE

INTRODUCTION

CHAPTER ONE : INTRODUCTION.

The aim of this thesis is to examine in detail the sedimentary processes which were active during the deposition of the Lower and Middle Cambrian rocks of the Welsh Basin. Rocks of this age outcrop in three main areas: the Harlech Dome, St Tudwal's Peninsula and Arfon. Sections were logged and compared vertically within the stratigraphic succession and laterally between these three areas.

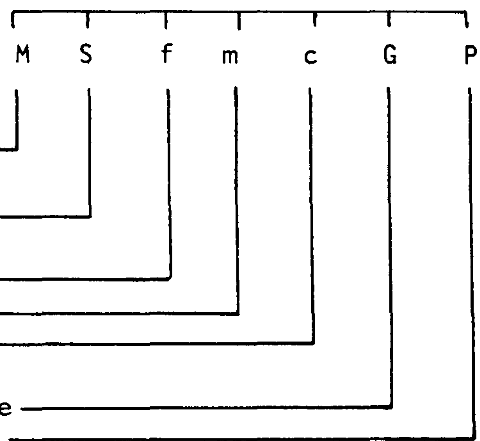
In order to construct a palaeogeographic model it is necessary to put sedimentation into a regional context (Chapter 2) and justify litho- and biostratigraphic correlation (Chapter 3). The facies and other aspects of the sedimentology are then described in detail for each of the three outcrop areas (Chapters 4, 5 and 6). The petrography is also described (Chapter 7) in order to shed light on possible source areas from which the Lower and Middle Cambrian sediments were derived. Finally the different strands of evidence are brought together in the last chapter (Chapter 8), general conclusions are made and palaeogeographic models attempted.

Much of the sedimentological data are presented as graphic logs, showing bed thicknesses, grain sizes and sedimentary structures (for key see Fig 1.1). Palaeocurrent data are also presented and where necessary they have been rotated back to the horizontal with respect to the plunge of folds and the dip of bedding. Field sketches, photographs, graphs, petrological point count data and some results from microprobe and stable isotope analyses are also presented.

Reference is also made to different aspects of turbidite research and some recent terminology is used in order to aid interpretation. Classical turbidites show Bouma (1962) sequences (Fig 1.2) either as complete or incomplete sequences (Walker 1965). However, rudaceous and some arenaceous sediments may show slightly different vertical sequences (Fig 1.3; Lowe 1982). Facies may also be defined (Fig 1.4; Mutti 1979) on the basis of sandstone bed

FIG 1.1 Key for Graphic Logs.

GRAIN SIZE (using Wentworth Scale)



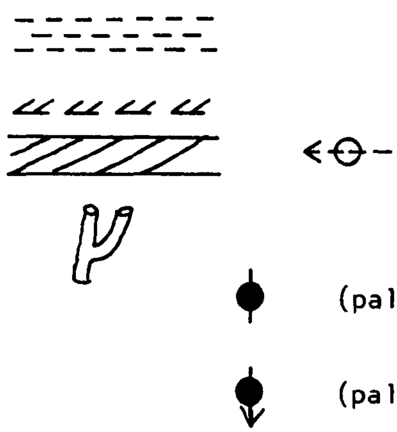
Mudstone

Siltstone

Sandstone: fine
medium
coarse

Conglomerate: granule
pebble

SEDIMENTARY STRUCTURES



Parallel Lamination:

Cross lamination:

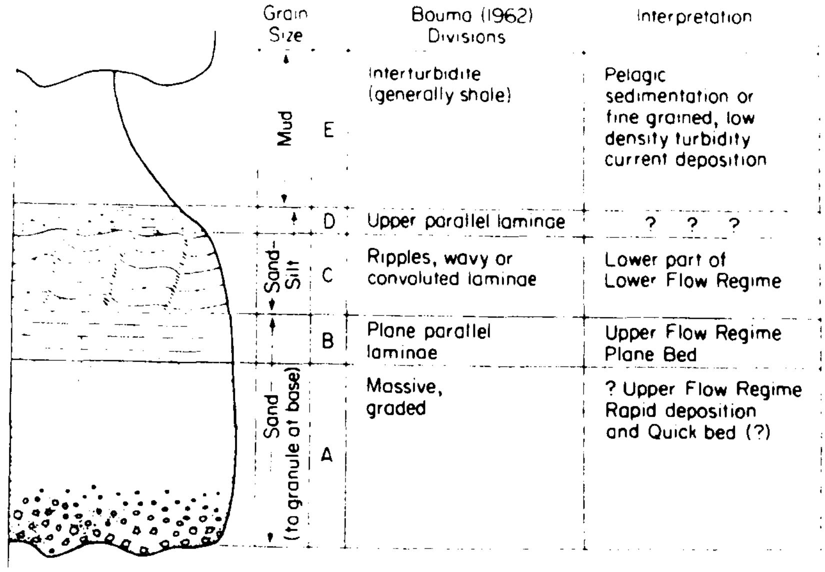
Cross-bedding:

Trace Fossils:

Groove casts/ scours: (palaeocurrent)

Flute casts: (palaeocurrent)

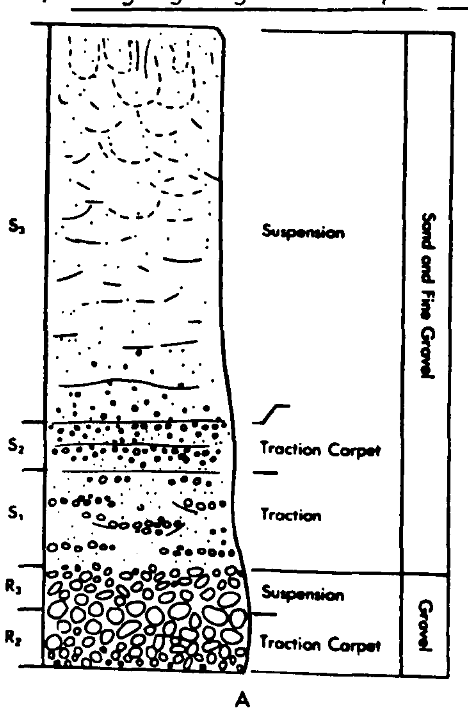
FIG 1.2 The Bouma (1962) Sequence.

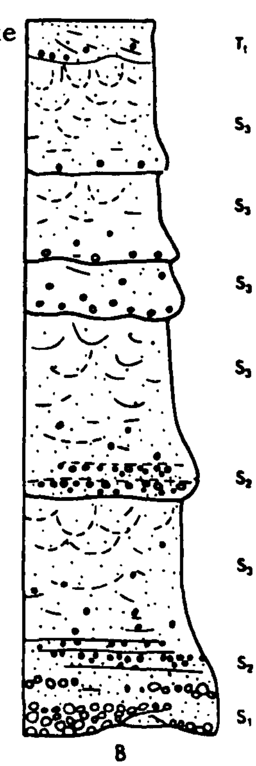


(from Middleton & Hampton 1976)

FIG 1.3 A) An ideal high density turbidite (Lowe 1982). Because of downslope separation of gravel and sand complete sequences such as this are rare.

B) Surging high density turbidite





(from Lowe 1982)

FIG 1.4 Turbidite Facies. (from Mutti 1979)

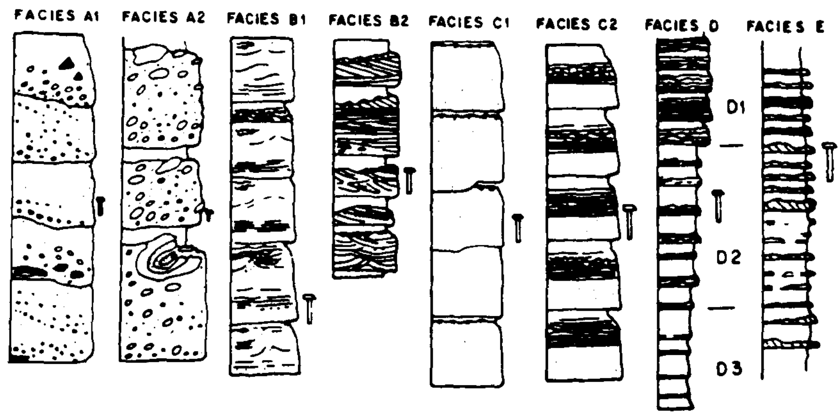
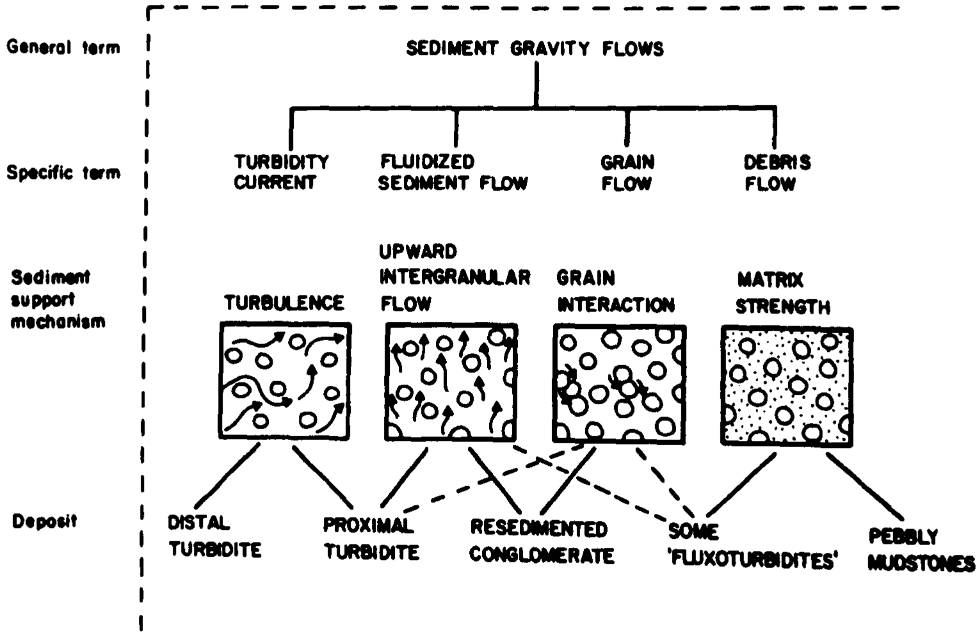


FIG 1.5 The different types of Sediment Gravity Flow and their sediment support mechanisms.

(from Middleton & Hampton 1976)



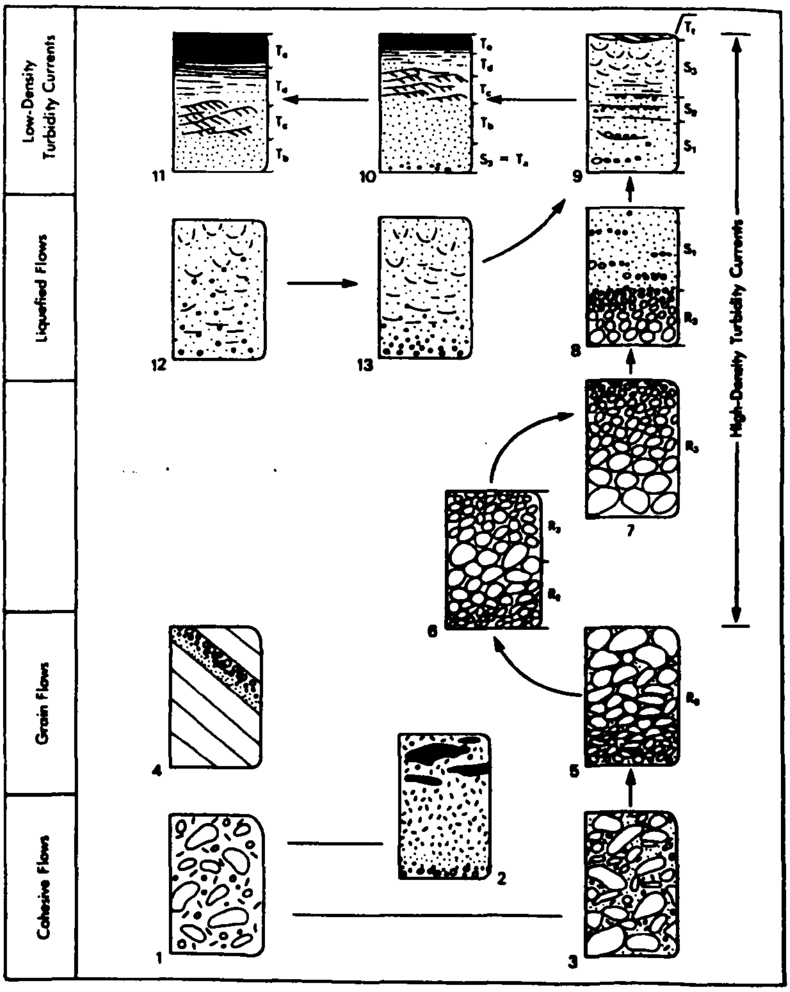
thickness and sedimentary structures.

The types of sedimentary structures that occur are related to the density of the flow from which they were produced. Dilute low density flows support grains within a turbulent suspension which is capable of transporting sediment for large distances by the process of autosuspension (Pantin 1979) and tend to produce sequences similar to the Bouma sequence. However studies of massive thick bedded sandstones (Stauffer 1967, Fisher 1971, Lowe 1982) indicate that they are probably deposited from high density flows where other grain support mechanisms are important (Fig 1.5; Middleton & Hampton 1976) and lead to a variety of types of deposits (Fig 1.6; Lowe 1976, 1982). Many of these flows are referred to by Lowe (1982) as high density turbidity currents. Grains may however have a complex transportational history since flows may undergo transformation to different types of flow (Fisher 1983).

Turbidites from the Lower and Middle Cambrian rocks of North Wales are also compared with the different fan models e.g. Mutti & Ricci Lucchi 1972, Walker 1978, Surlyk 1978.

FIG 1.6 The main types of deposit formed from Sediment Gravity Flows.

(from Lowe 1982)



CHAPTER TWO

REGIONAL SETTING

CHAPTER TWO : REGIONAL SETTING.

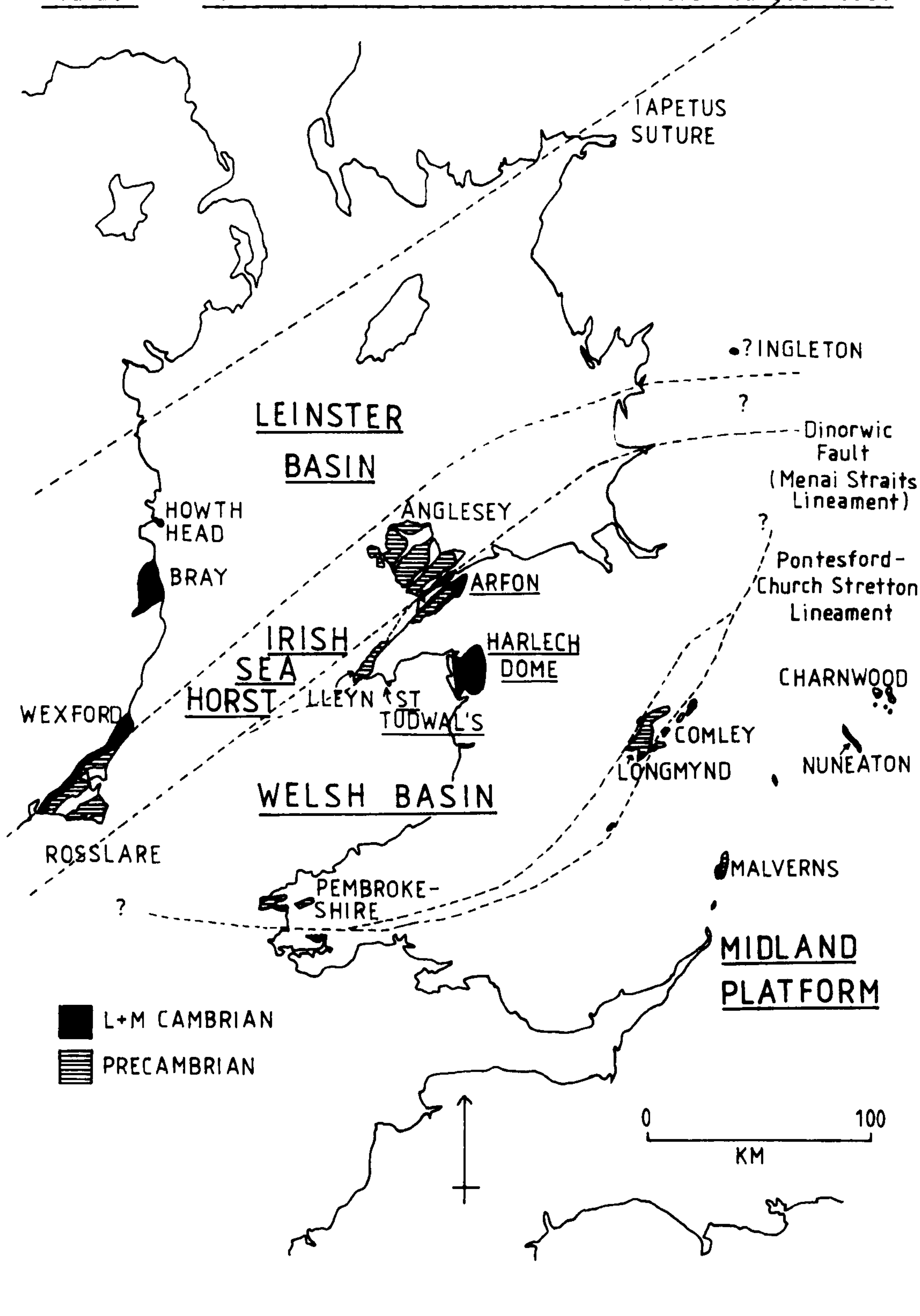
The Caledonides reflect orogenic activity ranging in age from the late Precambrian to Devonian and form a belt running through north-western Scandinavia, eastern Greenland, the British Isles, and eastern North America. The Caledonides of Britain are mainly exposed in Scotland, north-west and south-east Ireland, Wales and north-west England. The most important structural line within the British Caledonides is the Iapetus Suture, a north-east, south-west trending lineament. This lineament divides two areas which are underlain by basement of differing type and age (Brewer *et al.* 1983; Thorpe *et al.* 1984), which belonged to different faunal provinces prior to the Devonian (Wilson 1966) and had contrasting geological histories (Anderton *et al.* 1979). There is therefore evidence for spatial separation of the areas to the north-west and south-east of the suture. Possible remnants of oceanic crust (e.g. the Ballantrae Complex) and the occurrence of a Lower Palaeozoic accretionary prism in the Southern Uplands which was situated above a north-westward dipping subduction zone (McKerrow *et al.* 1977) suggest that Britain north-west and south-east of Iapetus Suture were separated by an ocean (Iapetus Ocean). The final closure of this ocean in the Devonian ended the Caledonide orogenic cycle. The Iapetus Suture is regarded as a former plate boundary between the Laurentian Plate to the north-west and Cadomian Plate to the south-east (Soper & Hutton 1984).

The British Caledonides south of Iapetus Suture can be divided into the following zones (Fig 2.1; Phillips *et al.* 1976):

1) The Leinster Basin.
2) The Irish Sea Horst.
3) The Welsh Basin.
4) The Midland Platform.

The Lower Palaeozoic rocks of the Leinster Basin and Welsh Basin are dominated by turbidite deposits in the

FIG 2.1 Tectonic Zones of the southern British Caledonides.



Cambrian and Silurian and by volcanics in the Ordovician. The thick sequences preserved in these areas of subsidence contrast with the Irish Sea Horst which was an area of non-deposition or erosion during most of the Lower Palaeozoic (Jones 1938). The Midland Platform, located to the east of the Welsh Basin, was an area of predominantly stable shelf deposition.

The Menai Straits Fault System (including the Dinorwic, Aber-Dinlle and Berw faults) and the Church Stretton-Pontesford Lineaments formed the north-western and south-eastern margins, respectively, to the Welsh Basin. In the Cambrian the Menai Straits Fault System was active during sedimentation (Reedman *et al.* 1984). The south-eastern margin of the Welsh Basin is poorly constrained at this time; no Lower and Middle Cambrian rocks are exposed between the Harlech Dome and Comley in Shropshire, so the precise position of the basin margin is not known.

In order to determine the framework in which the Lower and Middle Cambrian strata of North Wales were deposited it is necessary to examine the events which preceded their deposition. No Precambrian basement is exposed within the Welsh Basin, so our knowledge of late Precambrian events is limited to the surrounding zones, namely the Irish Sea Horst and Midland Platform.

Irish Sea Horst.

Precambrian rocks of the Irish Sea Horst are exposed on Anglesey and the Western Lleyn in Wales and County Wexford in south-east Ireland. The oldest rocks of the Irish Sea Horst occur within the Rosslare Complex of Wexford; they include gneisses which may have suffered deformation around 1600 Ma (Max 1975). Gneisses also outcrop on Anglesey and the Lleyn, though the relationship between these gneisses and the rest of the Mona Complex (the late Precambrian strata exposed on Anglesey, as defined by Greenly 1919) is unclear. Some workers (Greenly 1919; Barber & Max 1979) suggest the gneisses on Anglesey are ancient (i.e. early Proterozoic, Rosslare-type), high grade metamorphic basement, while others (Shackleton 1956, 1975) argue that the gneisses represent highly metamorphosed Mona Complex. Monian gneisses have yielded young isotopic ages (e.g. 595 $\pm$ 12 Ma for the Holland Arms gneiss; Beckinsale & Thorpe 1979) which are similar to the age of metamorphism in the rest of the Mona Complex. It is unclear whether the gneisses are yielding original metamorphic ages or the ages have been reset by later events. Therefore correlation between the Monian gneisses and the Rosslare Complex is a matter of debate and the areal extent of early Proterozoic basement is uncertain.

Wright (1977) correlated the Rosslare Complex with the Pentevrian of Brittany and suggested that ancient basement underlies most of the Caledonides of southern Britain. However no rocks of equivalent age are exposed between Rosslare and Brittany. An alternative view is that the basement to the Lower Palaeozoic rocks of southern Britain is relatively young (c. 900-400 Ma, Thorpe *et al.* 1984) and that ancient basement is restricted to the Irish Sea Horst where it occurs only as narrow fault-bounded slices (Gibbons 1983a).

The Mona Complex of Anglesey and the Lleyn Peninsula have been correlated with the Cullenstown Formation of south-east Ireland (Barber *et al.* 1981). The age of the Mona Complex is a matter of debate. Most structural and radiometric data (for a review see Gibbons 1984) indicate that the main period of Monian deformation occurred in the late Precambrian or early Cambrian (around 600 Ma), though some parts of the Mona Complex may have been deformed in the Lower Palaeozoic (Barber & Max 1979).

The occurrence in the Mona Complex of sedimentary melanges, basaltic pillow lavas, ultrabasic intrusions (ophiolitic interpretation see Thorpe 1972, 1974; non-ophiolitic interpretation see Maltman 1975) and blueschists, associated with calc-alkaline volcanics to the south-east of Anglesey suggest south-eastward dipping subduction of oceanic crust beneath Anglesey in late Precambrian times (Wood 1974; Thorpe *et al.* 1984). Gibbons (1983b) interpreted the Penmynydd Zone of south-eastern Anglesey (including the blueschists) as a ductile shear belt produced by strike-slip deformation between a low grade Mona Complex terrane and a high grade terrane composed of gneisses and granites. Thus the extent to which the Mona Complex was influenced by orthogonal subduction/collision or by strike-slip tectonics during the Precambrian is not well understood. Monian deformation occurred in late Precambrian to early Cambrian times, predating the Bwlch Gwyn Tuff of Anglesey and thus also predating its lateral equivalents in the Arfon Group which form the basal part of the Cambrian succession in mainland Wales. Monian deformation occurred at a similar time to the Cadomian orogeny (Cogne & Wright 1980), though it is unclear to what extent these two events are related. The Mona Complex is not exposed in the Welsh Basin or the Midland Platform and therefore its south-eastern extent is not known. Gravity data, however, indicates that Monian-type basement is present beneath Arfon (Reedman *et al.* 1984).

Nutt & Smith (1981) suggest that there was strike-slip movement on the Menai Straits Lineament during the Lower Palaeozoic, so that the Mona Complex of Anglesey may not have been adjacent to the Welsh Basin in Cambrian times. However the correlation of the Bwlch Gwyn Tuff, Careg-onen Beds and Baron Hill Beds of Anglesey with the Arfon Group of Arfon (Reedman *et al.* 1984) and the occurrence of clasts of Mona Complex type in the Cambrian sediments of mainland Wales (Greenly in Nicholas 1915; Gibbons 1983b, 1984) indicate that there was little strike-slip movement on the Menai Straits Lineament after the early Cambrian.

<u>Midland Platform.</u>

The Midland Platform shows three main phases of development in the late Precambrian (Thorpe *et al.* 1984):

1) The earliest phase, c.700-630 Ma: mainly represented by igneous complexes including the Malvernian Complex (Worcestershire), Johnson Complex (Dyfed), Stanner-Hanter Complex (Welsh Borders) as well as the Pebidian volcanics (Dyfed). The intrusives have volcanic arc characteristics and may be coeval with the Mona Complex of Anglesey.

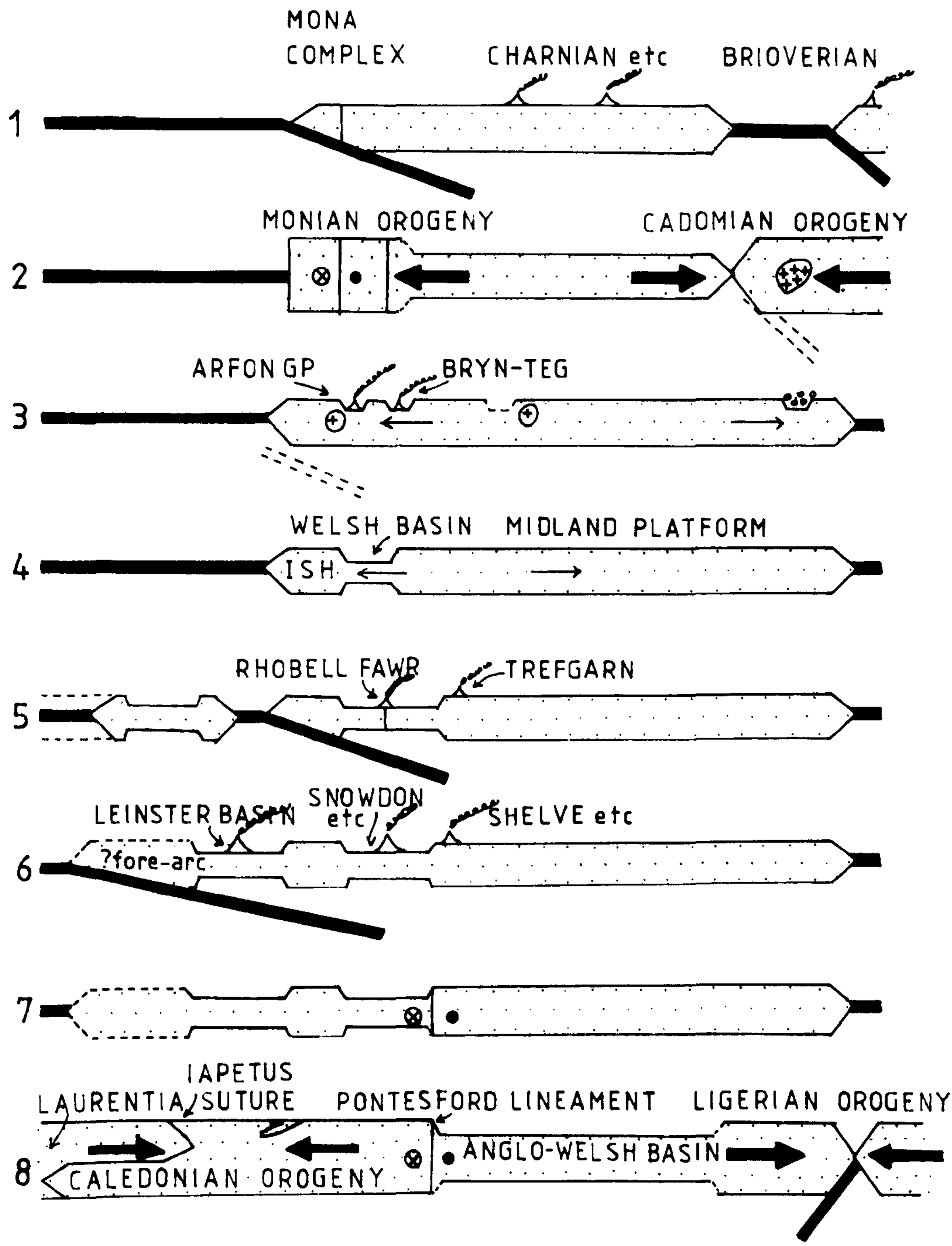
2) The middle phase, c.600-550 Ma: calc-alkaline volcanics and associated sediments including the Uriconian-Longmyndian (Shropshire), Caldecote Volcanics (Warwickshire) and Charnian (Leicestershire). They represent a later period of island arc volcanism.

3) The end Precambrian folding event with associated deformation and intrusions: e.g. St David's granophyre of Dyfed (between 650 and 570 Ma, Patchett & Jocelyn 1979), Ercall granophyre of Shropshire (533 $\pm$13 Ma, Patchett *et al.* 1980). The folding may be of similar age to deformation in the Monian of Anglesey and Cadomian of Brittany and may be related. The Sarn Granite of the Lleyn Peninsula is also of similar age (549 $\pm$19 Ma, Beckinsale *et al.* 1984).

The crust of the Midland Platform may be relatively young and have formed by the accretion of island arcs throughout much of the late Precambrian (Thorpe *et al.* 1984). Precambrian lineaments were probably reactivated in the Lower Palaeozoic and became major controls on sedimentation and the style of later deformation (e.g the Rhobell Fracture, Kokelaar 1979). Woodcock (1984 a, b) has shown that faults on the south-eastern margin of the Welsh Basin (in particular the Pontesford Lineament) had strike-slip displacements during the late Ordovician and there is evidence to suggest that there may have been late Precambrian movement on the Church Stretton Fault. However the amount of lateral movement between the Midland Platform and the Welsh Basin is poorly constrained.

In summary the following stages in the development of the Caledonides of southern Britain can be tentatively recognised (Fig 2.2):

FIG 2.2 Hypothetical evolution of the southern British Caledonides.



KEY

⊗ ● Strike-slip faulting

⊕ Granitic intusions

Volcanoes

ISH : Irish Sea Horst

1) Late Precambrian Subduction.

South-eastward dipping subduction occurred to the north-west of Anglesey from c.700 to 600 Ma (Dewey 1969; Thorpe *et al.* 1984). Thus the Mona Complex was probably deposited in a near-trench environment. However there are many different models; Baker (1973) infers a north-westward dipping subduction zone south of Anglesey, while Rast *et al.* (1976) suggest a passive margin setting for the Monian of Anglesey.

2) End Precambrian Collision.

c.600-550 Ma there is evidence for collisional events on the north-western and south-eastern margins of Cadomia (as defined by Soper & Hutton 1984). On the north-western margin collision occurred as part of a deep-rooted, strike-slip dominated orogeny (Gibbons 1983a). In Brittany collision may have been more orthogonal between Cadomia and a "Proto-African" plate (Anderton *et al.* 1979) or possibly between Armorica and southern Britain, resulting from the closure of an ocean along a south-eastward dipping subduction zone in the English Channel (Cogne & Wright 1980). The collision produced the Cadomian orogeny and the formation of the Cadomian Plate.

3) Early Cambrian Vulcanicity.

c.550 Ma either subduction recommenced or post-Cadomian tension allowed crustal melts to rise. Intrusions include the Ercall granophyre of the Midland Platform (which is unconformably overlain by Lower Cambrian sediments, Cope & Gibbons 1987), Sarn granite (Irish Sea Horst) and Twt Hill granite (Welsh Basin). The Arfon Group and Bryn-teg arc volcanics were erupted at this time in the Welsh Basin. The development of the Arfon basin in North Wales was probably, at least initially, volcano-tectonically controlled (Reedman *et al.* 1984).

4) Cambrian Sedimentation.

Turbidite deposition dominates Cambrian sequences in the

Welsh Basin, while shelf-dominated sequences occur in the Midland Platform. In the late Cambrian sedimentation exceeded subsidence and sequences show increasing evidence for deposition in shallow water (Crimes 1970a).

The early Cambrian of the Welsh Basin and Midland Platform have transgressive sequences which may be associated with the splitting up of a Proterozoic supercontinent (Piper 1982) in the late Precambrian. There is strong evidence that Cadomia and northern Britain were widely separated by Iapetus Ocean in the Cambrian, e.g. different faunal provinces (Cowie 1971). Palaeomagnetic data suggest that Cadomia may have been adjacent to the African-Arabian part of this supercontinent during the Precambrian and split off in the late Precambrian to early Cambrian (Thorpe *et al.* 1984). This rifting event may account for crustal tension in the Cambrian which led to early Cambrian development of localised basins (e.g. the Arfon Basin, Reedman *et al.* 1984) followed by more widespread subsidence in an essentially non-volcanic, passive margin setting. No evidence has yet been found for transtensional pull-apart basins analogous to Woodcock's (1984 a,b) interpretation of the upper Ordovician Welsh Basin.

5) <u>Tremadocian Vulcanicity and Uplift.</u>

The Rhobell Fawr volcanics of the eastern Harlech Dome (Kokelaar 1977) and Trefgarn volcanics of Dyfed indicate an island arc setting for the Welsh Basin in Tremadocian times. The island arc was produced by a south-eastward dipping subduction zone near the north-west margin of the Irish Sea Horst (Kokelaar *et al.* 1984) and some of the vulcanicity was concentrated on pre-existing Precambrian fractures, e.g. the Rhobell fracture (Kokelaar 1979).

Open north-south folding and uplift in late Tremadoc times (Roberts 1979) may have resulted from a mild collisional event which ended subduction, though the effects were very variable throughout North Wales. In parts of Wales Arenig strata lie unconformably on older rocks (Wells 1925) ranging in age from Precambrian to Tremadoc (Shackleton 1953). There was no pre-Arenig folding event in the Leinster Basin so it may have been separated from the Welsh Basin at this time.

6) Ordovician Subduction.

Bimodal basic and silicic volcanism in the Ordovician indicate that the Welsh Basin was a marginal (back-arc) basin occurring above a south-eastward dipping subduction zone (Kokelaar *et al.* 1984). The trench lay in the region of, or to the north-west of the Iapetus Suture (i.e. to the north-west of Tremadocian subduction) and subduction of oceanic crust was related to the gradual closure of Iapetus ocean. Sequences in the Leinster Basin are dominated by arc volcanics (Stillman 1986) and it is possible that part of the forearc was subducted beneath the Southern Uplands during late Silurian to early Devonian collision (Leggett *et al.* 1983). Stable shelf sedimentation predominated on the Midland Platform. The development of the southern Caledonides during the Ordovician is summarised by Anderton *et al.* (1979).

7) Silurian Sedimentation.

In contrast to the Ordovician, volcanism was uncommon in the Silurian, apart from the Skomer alkali basaltic volcanics (Dyfed). Turbidite deposition dominated in the Welsh Basin while shelf deposition persisted on the Midland Platform.

8) Siluro-Devonian Collision.

Soper & Hutton (1984) show that north-south compression (producing sinistral strike-slip on north-east south-west oriented faults) was produced by the collision of the Cadomian plate with Laurentia-Baltica (Cocks & Fortey 1982) in the Siluro-Devonian. Collision resulted in deformation, metamorphism and igneous intrusion within the Caledonide belt (Watson 1984). Uplift associated with this orogenic activity resulted in the shedding of sediment into Devonian molasse basins.

In Armorica a separate collisional event occurred- the Ligerian Orogeny which was also Siluro-Devonian in age (Cogne & Lefort 1985). This was associated with a north-westward dipping subduction zone which lay to the south-east of Brittany (Leeder 1982).

CHAPTER THREE

STRATIGRAPHY

CHAPTER THREE : STRATIGRAPHY.

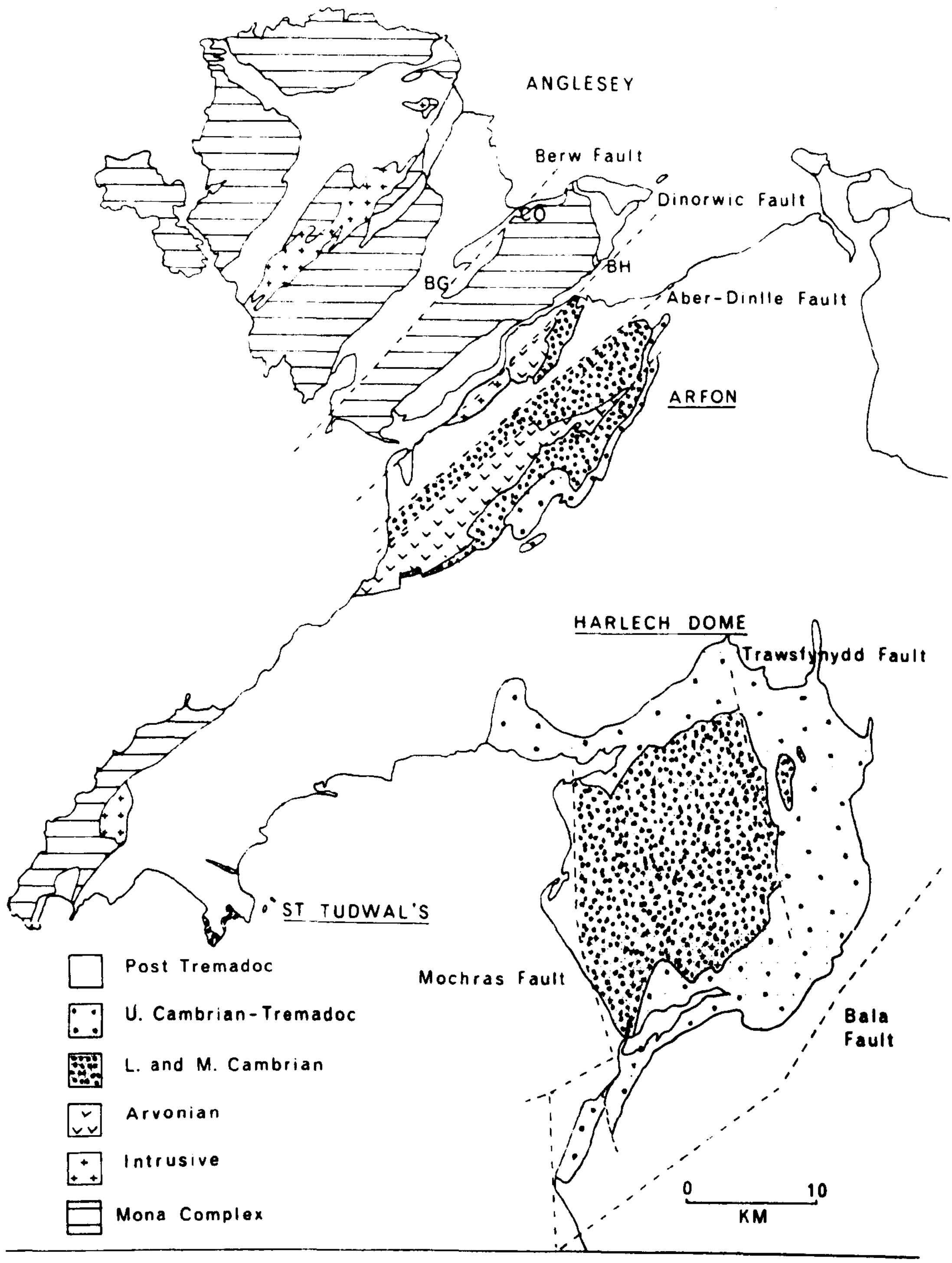
The Lower and Middle Cambrian of North Wales outcrop in three main areas: the Harlech Dome, St Tudwal's Peninsula and Arfon (Fig 3.1), occurring within the confines of the Lower Palaeozoic Welsh Basin (Fig 2.1), as defined by Jones (1938). Small, isolated outcrops of strata of possible Lower Cambrian age also occur on Anglesey (Reedman *et al.* 1984).

Cambrian sedimentation in the Welsh Basin began with the deposition of a thick sequence of volcanics and volcaniclastics which are generally acidic in the Arfon region (Arfon Group) and basic and intermediate in the Harlech Dome (Bryn-teg Volcanic Formation). These volcanics probably overlie a previously metamorphosed basement of Monian type (as exposed on Anglesey and the western Lleyn) or an intrusive-dominated basement of Sarn Complex type (Reedman *et al.* 1984; Gibbons 1983a). In the Harlech Dome the volcanics are overlain by conglomerates, sandstones and siltstones (Dolwen Formation) which were deposited mainly in shallow water (Allen & Jackson 1978). This formation was followed by deeper water deposition (represented by slates of the Llanbedr Formation). In contrast, the early Cambrian succession of the Arfon region is dominated by volcanics and volcanic-derived sediments (Minffordd and Bangor Formations north-west of, and the Fachwen Formation south-east of the Aber-Dinlle Fault). Remnants of similar sequences also occur on Anglesey. The Bangor and Fachwen Formations of Arfon pass upwards into a much thicker sequence of slates (the Llanberis Slates Formation) than is found in the Harlech Dome.

The upper part of the Lower and Middle Cambrian succession contains alternating sequences of:

a) sand-rich turbidites: the Rhinog and Barmouth Formations of the Harlech Dome, the Hell's Mouth Grits and Cilan Grits of St Tudwal's and the Bronllwyd Grit Formation of Arfon.

FIG 3.1 Outcrop map of the Precambrian and Cambrian rocks of North Wales.



b) silt/mud-rich turbidites: the Hafotty and Gamlan Formations of the Harlech Dome, the Mulfran Beds and Caered Flags and Mudstones of St Tudwal's and the Llanberis Slates Formation of Arfon. A manganiferous horizon occurs within the Hafotty Formation of the Harlech Dome and Mulfran Beds of St Tudwal's.

The correlation between sequences in the Harlech Dome, St Tudwal's, Arfon and Anglesey (Figs 3.2, 3.3) is reviewed in the following sections.

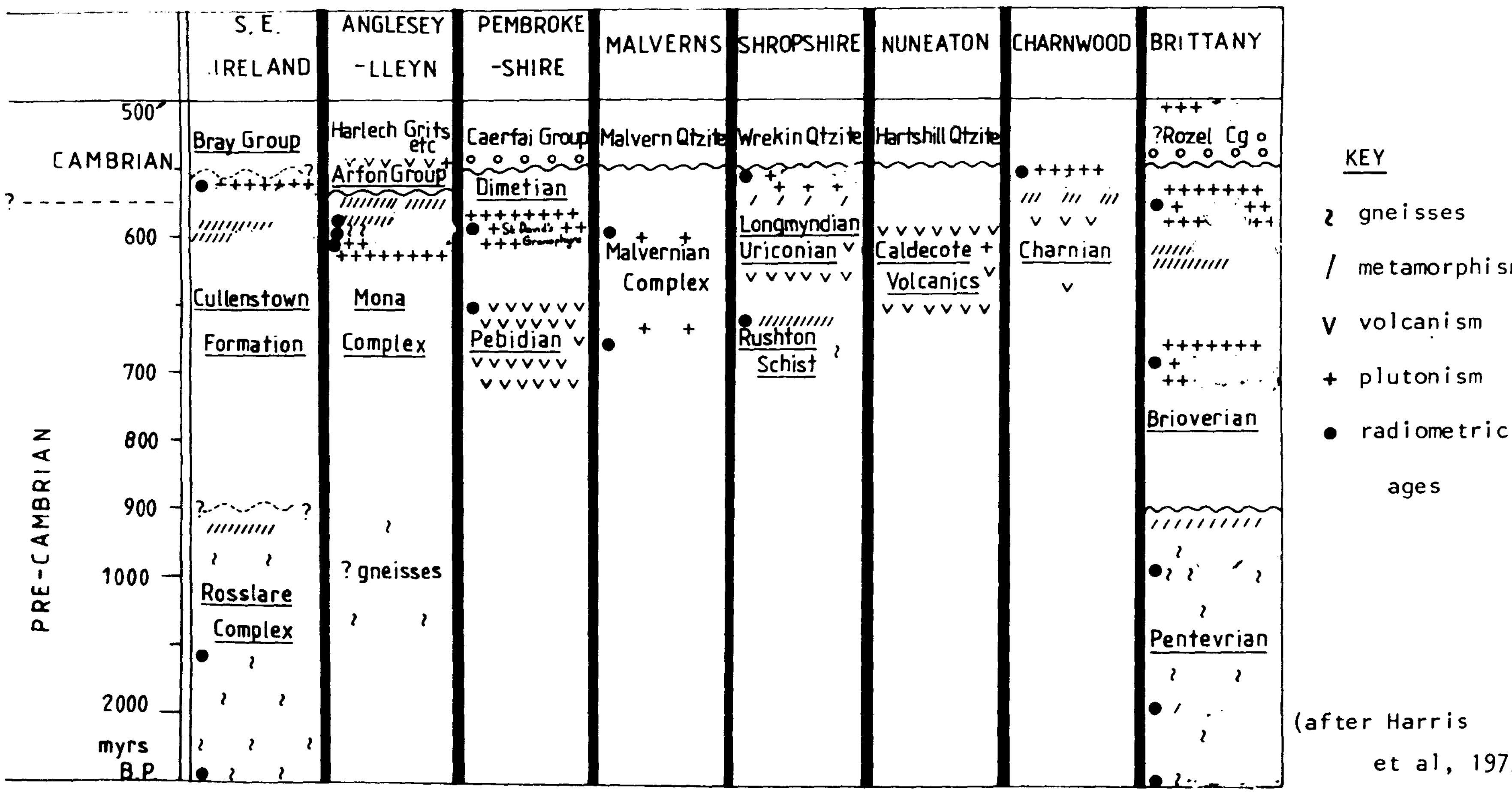
1) Harlech Dome.

The Bryn-teg Volcanic Formation. The Bryn-teg borehole proved the existence of basic and intermediate lavas and volcaniclastics below the lowest exposed Cambrian sediments of the Harlech Dome. At the base of the overlying Dolwen Formation, conglomerates contain abundant silicic volcanic fragments and indicate a markedly different provenance to the underlying Bryn-teg Volcanic Formation. There is no angular unconformity at the contact of the two formations but the abrupt change in source may indicate a disconformable relationship (Allen & Jackson 1978). Whether the Bryn-teg Volcanic Formation in the Harlech Dome correlates with Arfon Group in the Arfon region is unclear; the former is generally less acidic than the latter indicating that they probably represent separate volcanic episodes.

The HARLECH GRITS GROUP overlies the Bryn-teg Volcanic Formation and is a conformable sequence of clastic deposits divided into the following formations:

a) The Dolwen Formation is comprised of conglomerates near the base of the succession passing up into sandstones and siltstones. Allen & Jackson (1978) tentatively correlate the Upper Sandstones and Siltstones divisions of the Dolwen Formation with sandstones and slates within the Llanberis Slates Formation of Arfon. However, in detail these sequences are quite dissimilar. The presence of

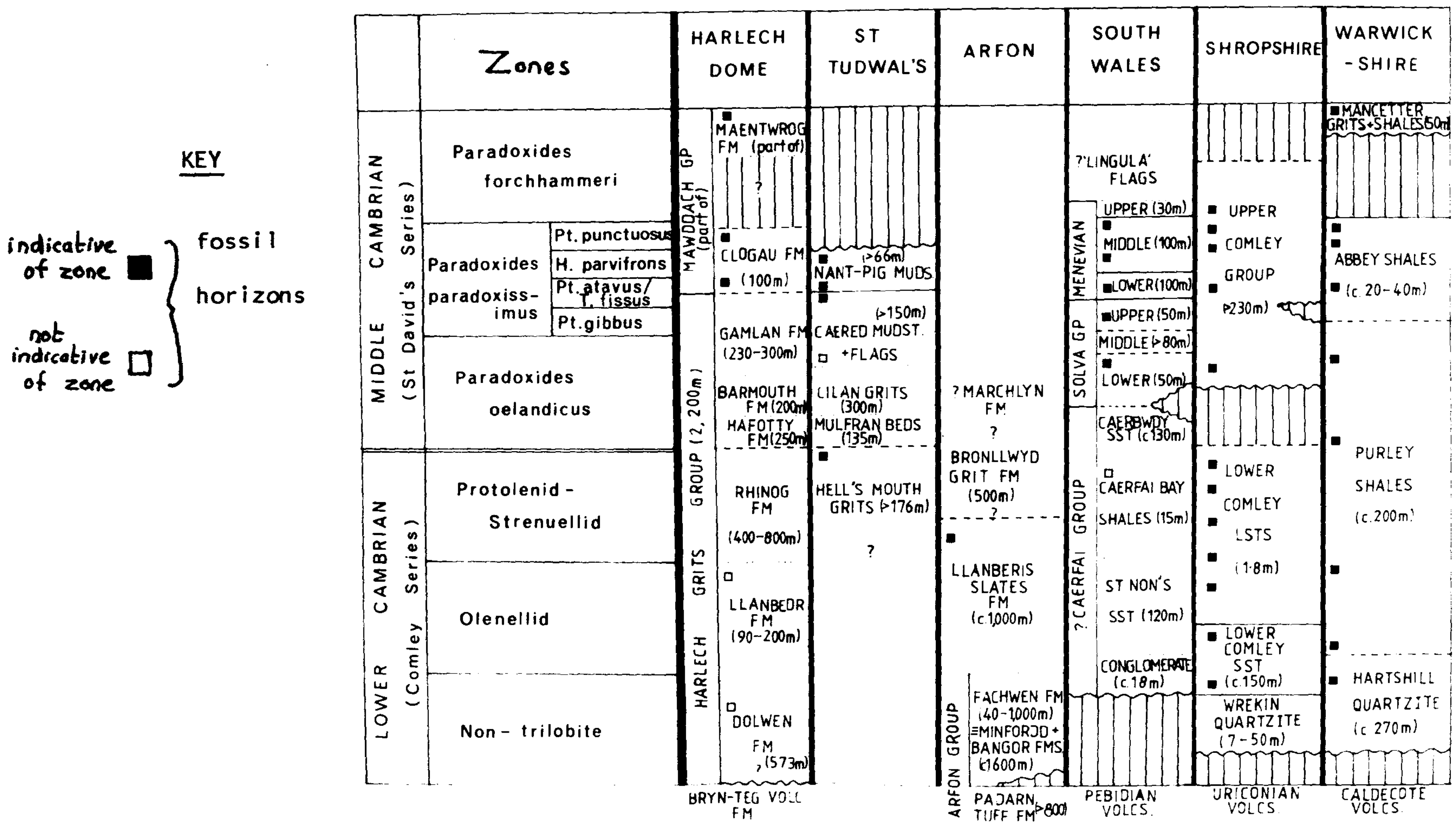
FIG 3.2 Late Precambrian Correlation Chart, south of Iapetus Suture.



(after Harris et al, 1975)

FIG 3.3 Lower and Middle Cambrian Correlation Chart for southern Britain.

(after Cowie *et al*. 1972)



Platysolenites, a silicified worm burrow suggests that the Dolwen Formation is Lower Cambrian in age (Rushton 1978).

b) The Llanbedr Formation is composed mainly of slates. Lingulid brachiopods have been found near Llyn Cwm Mynach which confirm that these rocks are younger than Precambrian (Lockley & Wilcox 1978). The report of horny brachiopods cf. *Kutorgina* found "close to the Harlech golf links" in beds which "underlie the thick false-bedded grits of Harlech Castle" (Fearnsides 1910a) has not been confirmed (Rushton 1974).

c) The Rhinog Formation consists predominantly of thick bedded sandstones, and although body fossils are absent, they contain a relatively diverse trace fossil fauna including *Planolites, Phycodes* and *Arenicolites.*

d) The Hafotty Formation is unfossiliferous but near its base contains a laterally persistent manganiferous horizon. The Manganese Bed can be correlated throughout most of the Harlech Dome as well as with the Manganese Shales (Mulfran Beds) of St Tudwal's Peninsula (Nicholas 1915; Matley & Wilson 1946) and possibly even with a manganese ore bed in the Chamberlain's Brook Formation of south-east Newfoundland (Mohr & Allen 1965). Trilobites in the upper part of the Hell's Mouth Grits of St Tudwal's indicate a late Lower Cambrian age (Bassett *et al.* 1976). This suggests that the Mulfran Beds which almost directly overlie the fossiliferous horizon are also late Lower Cambrian or early Middle Cambrian in age as would be their lateral equivalent in the Harlech Dome- the Hafotty Formation.

e) The Barmouth Formation is comprised of thickly bedded sandstones and contains a few simple trace fossils.

f) The Gamlan Formation consists largely of thin bedded siltstones and sandstones and contains abundant trace fossils. Matley & Wilson (1946) recorded a trilobite fragment and a few inarticulate brachiopods from this formation, but no fossils useful for more detailed correlation were found.

The MAWDDACH GROUP conformably overlies the Harlech Grits Group. The lowest formation, the Clogau Formation, is a sequence of mainly dark, organic-rich and silty mudstones. They contain a trilobite fauna which is assigned to the middle St David's Series (Cowie *et al.* 1972; Allen *et al.* 1981). The Clogau Formation can be correlated on fossil evidence with the Nant-pig Mudstones of St Tudwal's Peninsula and the Lower Menevian of South Wales, which constrains the age of the Harlech Grits Group as pre-middle St David's Series.

2) St Tudwal's Peninsula.

The early Cambrian rocks of St Tudwal's are broadly similar to the Harlech Grits Group of the Harlech Dome. The succession on St Tudwal's is divided into the following units:

a) The Hell's Mouth Grits are the oldest rocks on St Tudwal's and the base of this unit is not exposed. They contain thick bedded sandstones intercalated with thinner bedded sandstones and mudstones. Trace fossils near the base of the exposed sequence, including *Phycodes*, indicate a probable Cambrian age. Trilobites, including *Hamatolenus (Myopsolenus) douglasi*, *Kerberodiscus succinctus*, and *Serrodiscus ctenoa*? (Bassett *et al.* 1976), occur near the top of the unit and indicate a late Lower Cambrian age (Protolenid-Strenuellid Zone, Comley Series). Potter (1974) confirmed this age on the basis of palynological evidence (zone S1A). Lithologically the Hell's Mouth Grits are similar to the Rhinog Formation of the Harlech Dome.

b) The Mulfran Beds contain manganiferous shales which can be correlated with the Hafotty Formation of the Harlech Dome and may be early Middle Cambrian in age.

c) The Cilan Grits contain a diverse trace fossil fauna (Crimes 1970a and own observations) and are early Middle Cambrian (St David's Series) in age on palynological evidence (zone S1B, Potter 1974). The Cilan Grits can be correlated with the Barmouth Formation of the Harlech Dome on the basis of similar lithology. However, the Cilan Grits are 100m thicker than the Barmouth Formation (without any comparable increase in bed thicknesses), which suggests that the unit's upper and/or lower boundaries may be diachronous.

d) The Caered Mudstones and Flags are divided into three units (Nicholas 1915):

i) The Lower Caered Mudstones. Matley *et al.* (1939) reported the presence of "*Agnostus*" 30m above the top of the Cilan Grits. However this discovery has not been repeated and precise identification has not been possible, though a Middle Cambrian age is indicated.

ii) The Caered Flags. Unfossiliferous.

iii) The Upper Caered Mudstones. They contain abundant fossils of the *Tomagnostus fissus* Zone of the St David's Series (Nicholas 1915, 1916; Rushton 1974). This assemblage is equivalent to the Lower Menevian of South Wales. Microflora also indicate a Middle Cambrian age (zone S1C; Potter 1974).

The Caered Mudstones and Flags are overlain by the Nant-pig Mudstones which contain a fauna diagnostic of the Middle Cambrian *Tomagnostus fissus* and *Hypagnostus parvifrons* Zones of the St David's Series (Nicholas 1915). The Nant-pig Mudstones are equivalent in age to the lower Clogau Formation of the Harlech Dome and the Middle Menevian of South Wales.

3) Arfon.

The ARFON GROUP (Reedman *et al.* 1984) is over 4000m thick and is exposed in two inliers in the Arfon region (the Bangor and Padarn ridges) which are separated by the north-east, south-west trending Aber-Dinlle Fault. The Arfon Group is divided into:

a) The Padarn Tuff Formation: a thick sequence of predominantly acidic ashflow tuffs which can be correlated directly with the Clogwyn Volcanics of south-west Arfon. The base is not seen but the formation probably rests unconformably on Monian-type basement by analogy with deposits of similar age on Anglesey (the Bwlch Gwyn "felsite"; Reedman *et al.* 1984). There was no large time gap between deposition of the Padarn Tuff Formation and the overlying formations (Wood 1969), which are Cambrian in age. The Padarn Tuff Formation is therefore either late Precambrian or more likely, early Lower Cambrian in age.

b) Different sequences overlie the Padarn Tuff Formation in different parts of Arfon:

i) The Minffordd Formation: volcanics and volcaniclastics, followed by the Bangor Formation: conglomerates and tuffites, outcrop on the Bangor Ridge (north-west of the Aber-Dinlle Fault). Sponge spicules from the Minffordd Formation indicate a post-Precambrian age for at least the upper part of the Arfon Group (Reedman *et al.* 1984). Since the Minffordd Formation occurs stratigraphically below the Llanberis Slates Formation, which contains Lower Cambrian trilobites (see below), a Lower Cambrian age is also accepted for the Minffordd and Bangor Formations.

ii) The Fachwen Formation: conglomerates, sandstones, siltstones and tuffs outcrop on the Padarn Ridge (south-east of the Aber-Dinlle Fault). The contact between the basal conglomerates of the Fachwen Formation and the Padarn Tuff was traditionally regarded as the base of the Cambrian (for a review see Wood 1969), though a lack of faunal control makes this difficult to prove or disprove. Wood (1969) showed that there is no pronounced unconformity or evidence for a significant time gap between the Padarn Tuff and the conglomerates of the Fachwen Formation, since vulcanicity continued in Fachwen Formation times and in places the contact between the two appears to be gradational.

iii) The Tryfan Grit Group, the Cilgwyn Conglomerate and the Glog Grit Group are the units which divide up this part of the succession in south-west Arfon (Morris and Fearnsides 1926).

The Arfon Group is overlain by:

a) The Llanberis Slates Formation: siltstones and mudstones with occasional turbidite sandstones. Some of the thicker sandstone units can be correlated throughout the region, e.g. the Red and Dwndwr Grits of Llanberis with the Dorothea and Pen-y-Bryn Grits respectively of Nantlle. Trilobites have been found at the top of the slates (within the "Green Slates") including *Pseudatops viola, Protolenus*?, *Strenuella*? and *Serrodiscus*? (Woodward 1888; Howell and Stubblefield 1950). This fauna has a stratigraphic range including both the Lower and Middle Cambrian. However, although the species *Pseudatops viola* has only been found in the Llanberis-Bethesda area, all other occurrences of the genus *Pseudatops* have been near the base of the Protolenid-Strenuellid Zone of the Comley Series. Since trilobites from St Tudwal's indicate an upper Protolenid-Strenuellid Zone age the Llanberis Slates Formation are tentatively interpreted as being slightly older (Wood 1969; Rushton 1974).

b) The Bronllwyd Grit Formation: a sequence of mainly sand-rich turbidites which can be correlated directly with the Cymffyrch Grit of south-west Arfon (Morris and Fearnsides 1926). There are several possibilities for the age of the Bronllwyd Grit Formation:

i) The Bronllwyd Grit Formation directly overlies the fossiliferous "Green Slate". If this contact is conformable (Crimes 1970a and personal observations), then the Bronllwyd Grit Formation is probably Lower Cambrian in age and can therefore be correlated with the Hell's Mouth Grits of St Tudwal's. The Bronllwyd Grit Formation appears to be overlain conformably by the Marchlyn Formation which is unfossiliferous in its lower part, though its upper part contains fossils which indicate an Upper Cambrian (Merioneth Series) age (Reedman *et al.* 1983). Alternatively if the Bronllwyd Grit Formation is of Lower Cambrian age then either there is a disconformity at the top of the formation or sedimentation was sufficiently slow during deposition of the Marchlyn Formation to span the entire Middle Cambrian and part of the Upper Cambrian. It is however possible that the Bronllwyd Grit Formation was deposited partly in the Lower and partly in the Middle Cambrian.

ii) Wood (1969) correlated a manganiferous bed in the Bronllwyd Grit Formation with the Manganese ore-bed (Hafotty Formation) of the Harlech Dome and the Manganese Shales (Mulfran Beds) of St Tudwal's. He therefore suggested that the Bronllwyd Grit Formation was early Middle Cambrian in age, which would require disconformities at the base and possibly at the top of the formation. However, manganiferous horizons are poorly developed in the Bronllwyd Grit Formation and also occur in the Gamlan Formation (upper Middle Cambrian) of the Harlech Dome.

iii) Since the Marchlyn Formation may be Upper Cambrian in its lower as well as its upper part and its lower boundary appears to be conformable. Morris & Fearnsides (1926) suggested that the Bronllwyd Grit Formation could be correlated with the Maentwrog Formation (Mawddach Group) of the Harlech Dome and therefore was uppermost Middle Cambrian to early Upper Cambrian in age; an Upper Cambrian age is also favoured by Howells *et al.* (1985).

Evidence for disconformities within the succession, however, is lacking. The sequence is therefore probably conformable. It is possible that the Bronllwyd Grit Formation is late Lower Cambrian in age and can be broadly correlated with the Rhinog Formation of the Harlech Dome and the Hell's Mouth Grits of St Tudwal's.

4) <u>Anglesey.</u>

The Arfon Group is present in three small outcrops on Anglesey: at Bwlch Gwyn (<u>BG</u> in Fig 3.1), Careg-onen (<u>CO</u>) and Baron Hill (<u>BH</u>). The Bwlch Gwyn "felsite" is correlated with the Padarn Tuff Formation on lithological grounds (Reedman *et al.* 1984). Sponge spicules were found at Careg-onen (Greenly 1919) and are probably of Cambrian age.

The Arfon Group on Anglesey is underlain by Mona Complex; the age of the latter has been a matter of controversy. The traditional view has been that the Mona Complex is entirely Precambrian (Greenly 1919). However on the basis of structural (Barber & Max 1979) and microfloral evidence (Muir *et al.* 1979) from the Gwna Group, parts of the Mona Complex may be Cambrian.

Correlation of the Lower and Middle Cambrian in North Wales.

A lack of fossils useful for correlation in the Lower and Middle Cambrian of North Wales prevents precise biostratigraphic correlation except in the upper parts of sequences i.e. above the Gamlan Formation in the Harlech Dome and above the Caered Flags in St Tudwal's, which are of late Middle Cambrian age and younger. Below these units the successions of the Harlech Dome and St Tudwal's are sufficiently similar lithologically to allow relatively close correlation. Correlation between these sequences and Arfon is much more difficult (Figs 3.2 and 3.3). The correlation scheme adopted is similar to that of Cowie *et al.* (1972) and is summarised in Fig 3.3. Thus two periods of sand-rich turbidite deposition are recognised:

i) Rhinog Formation - Hell's Mouth Grits - Bronllwyd Grit Formation (late Lower Cambrian).

ii) Barmouth Formation - Cilan Grits (early Middle Cambrian).

In general sediments deposited prior to the Rhinog Formation and its lateral equivalents are much more laterally variable and thus correlation is more difficult in this part of the sequence.

Correlation of the Lower and Middle Cambrian of North Wales with other areas.

a) Midland Platform (including South Wales).

Lower and Middle Cambrian sediments outcrop in the following areas of the Midland Platform:

1) South Wales. Precambrian Pebidian extrusives and Dimetian intrusives are overlain unconformably by the Caerfai Group (Cox *et al.* 1930), which is comprised of conglomerates, sandstones and shales deposited in shallow water. The Caerfai Group contains fossils which indicate a Cambrian or later age. However it must predate the overlying Lower Solva Beds which were deposited in the early Middle Cambrian. Therefore the Caerfai Group is probably Lower Cambrian in age. Higher in the sequence, the Solva and Menevian Groups contain a relatively complete development of Middle Cambrian (St David's Series) faunal zones. The upper part of the succession can be correlated with the Caered Mudstones and Nant-pig Mudstones of St Tudwal's and the Clogau Formation of the Harlech Dome.

2) Shropshire. Precambrian Uriconian volcanics are unconformably overlain by Lower Cambrian sandstones (Wrekin Quartzite). Overlying the Wrekin Quartzite is a condensed sequence consisting of:

i) The Lower Comley Sandstone (early Comley Series, Lower Cambrian).

ii) The Lower Comley Limestones which contain a rich and varied fauna (Cobbold 1927) from the Olenellid and Protolenid-Strenuellid Zones of the Comley Series.

iii) The Upper Comley Sandstones (St David's Series, Middle Cambrian) which overlie the Lower Comley Limestones with a contact varying from conformable to disconformable in different parts of the Comley area.

3) Nuneaton. Precambrian Caldecote volcanics are unconformably overlain by the Hartshill Quartzite which, in its upper part, contains an abundant small shelly fossil fauna indicative of the "Non- trilobite Zone" of the Comley

Series (Brasier *et al.* 1978). This is overlain by units rich in trilobites including the Purley Shales (upper Comley and lower St David's Series) and Abbey Shales, the latter of which can be correlated with the Menevian rocks of Wales (Rushton 1966).

4) Malvern Hills. Within the Malvern Hills the Malvern Quartzite and Hollybush Sandstone are exposed and both contain fossils of Comley Series age (Groom 1902).

5) Lickey Hills. The Lickey Quartzite is thought to be of similar age to the Malvern and Hartshill Quartzites on lithological grounds (Eastwood *et al.* 1925).

Most of the Cambrian sequences of the Midland Platform begin with mature, well-sorted, quartz-rich sandstones including the Wrekin, Hartshill, Lickey and Malvern Quartzites, deposited during the Lower Cambrian transgression (Brasier 1980, 1982; Brasier & Hewitt 1979). Above the quartzites, sequences are either mud-dominated (Nuneaton), or contain condensed sequences mainly of limestones and glauconitic sandstone (Shropshire). The sequences on the Midland Platform represent relatively slow, shallow-water deposition (in comparison to the Welsh Basin) on a continental shelf. Correlation with the Welsh Basin is possible on faunal grounds in the Middle Cambrian, but sequences are too dissimilar to allow detailed lithostratigraphic correlation.

b) Leinster Basin.

The Lower and Middle Cambrian rocks of the Leinster Basin outcrop in three main areas of south-east Ireland:

1) Wexford. The Bray Group is over 2,300m thick in Wexford and is comprised mainly of proximal turbidites with some distal turbidites (Shannon 1978). However, its base is not exposed. Within the equivalent Cahore Group, Lower Cambrian acritarchs have been found (Gardiner & Vanguestaine 1971), while higher in the sequence, trace fossils including *Oldhamia*, are indicative of a Middle

Cambrian age (Crimes & Crossley 1968). The Bray Group is overlain by the mudstone dominated Ribband Group, which contains acritarchs of late Middle to early Upper Cambrian age (Smith 1977), underlying beds containing Arenig and Llanvirn graptolites and trilobites (Brenchley *et al.* 1967).

2) Bray. The Bray Group here is over 6,500m thick, is comprised of proximal turbidites and includes quartzites of problematical origin (Shannon 1980). The upper part of the group contains acritarchs which indicate a late Lower to early Middle Cambrian age (Bruck *et al.* 1974).

3) Howth Head. The Cambrian succession here is quite different to the sequence at Bray. The lower part of the sequence (the Censure Group) is comprised of quartzites overlain by sandstones, siltstones and mudstones, some of which show large-scale slumping (van Lunsen and Max 1975). Trace fossils (Crimes 1976) and acritarchs suggest the Censure Group is middle Lower to early Middle Cambrian in age (Gardiner & Vanguestaine 1971). The Nose of Howth Group overlies the Censure Group and is comprised of olistostrome deposits with turbidites; it contains *Oldhamia*, which may indicate a Middle Cambrian age (Crimes 1976).

Evidence for deposition of Lower and Middle Cambrian sediments in the English part of the Leinster Basin is generally lacking. The Ingleton Group, a sequence of turbidites over 800m thick outcropping in the Pennines, were traditionally regarded as Precambrian in age. However O'Nions *et al.* (1973) obtained a dewatering date of 505 ± 7 Ma from the Ingleton Group which may suggest a possible Cambrian depositional age.

Accurate correlation is difficult within the Leinster Basin. The Bray Group and its lateral equivalents are probably late Lower Cambrian to early Middle Cambrian in age and thus represent proximal turbidite deposition at a similar time to Rhinog Grit deposition in the Welsh Basin. However the Lower and Middle Cambrian of the Welsh Basin differs from sediments of similar age in the Leinster Basin in the following ways:

i) Mud deposition was dominant in Middle to Upper Cambrian times (Ribband Group) in the Leinster Basin; there are no close analogues in the Welsh Basin.

ii) The Leinster Basin contains quartzite beds and sedimentary melanges, lithologies uncommon in the Welsh Basin.

iii) Slumping is more common in the Leinster Basin.

iv) Palaeocurrents and facies interpretations from the Leinster Basin (Shannon 1980), when compared with those for the Welsh basin, seem to indicate that deposition occurred in separate turbidite systems.

CHAPTER FOUR

HARLECH DOME

CHAPTER 4 : THE HARLECH DOME.

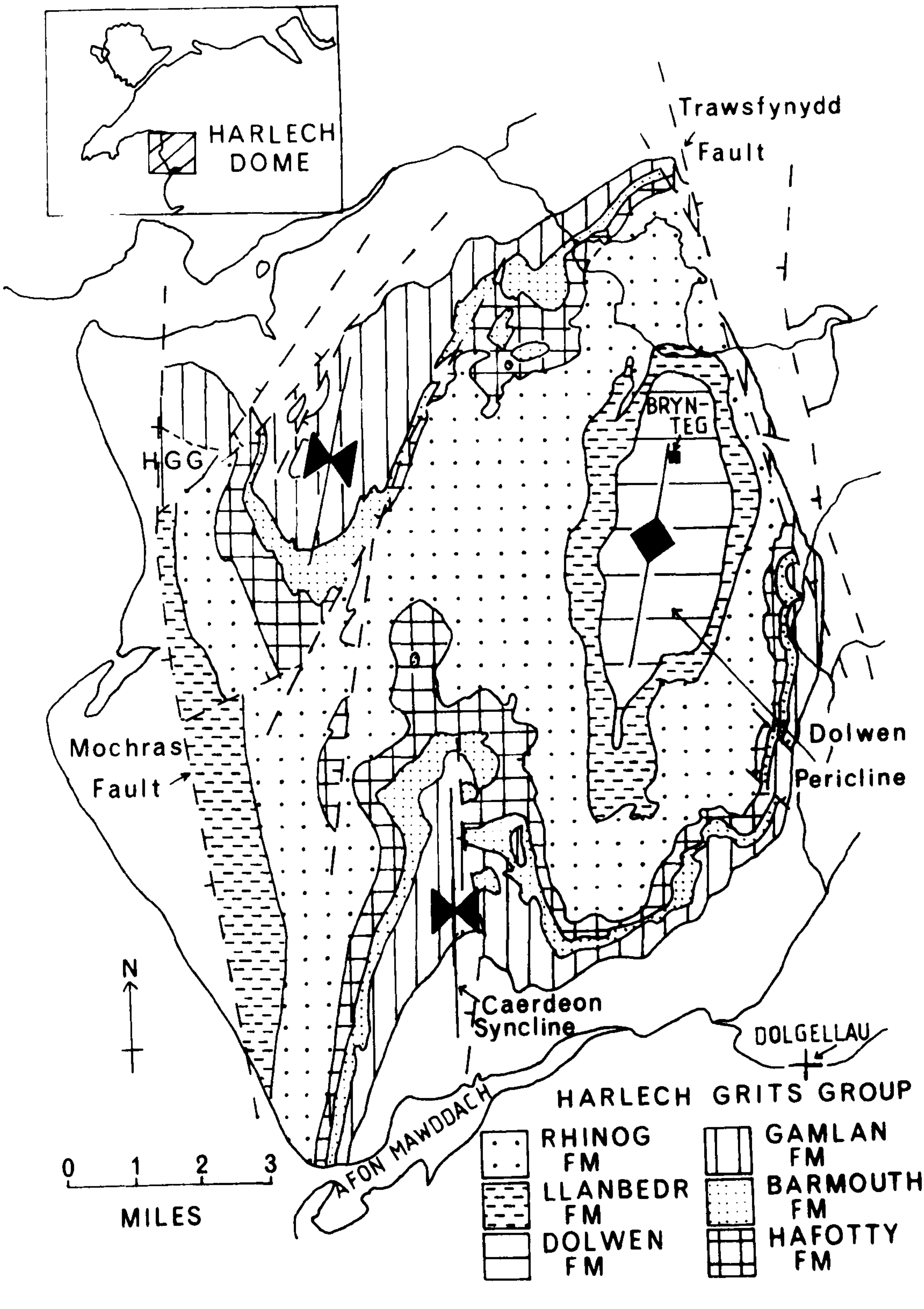
4.1 : Introduction.

The Harlech Dome is the area of North Wales to the east of Tremadoc bay, situated north of the Mawddach estuary and south of the Vale of Ffestiniog. Predominantly Cambrian rocks are exposed in this area, surrounded by Ordovician rocks on the Dome's northern, eastern and southern sides. The N-S striking Mochras Fault marks the sharp western boundary to the Dome. A thick Mesozoic and Tertiary succession was proved west of the Mochras Fault in the Mochras borehole (Woodland 1971).

The overall structure of the Dome is relatively simple; open folding predominates and fold axes strike N-S (Fig 4.1). Major folds include the Dolwen Pericline which exposes the oldest rocks of the area and the Caerdeon Syncline which plunges gently southwards. The dip of bedding is generally low (usually less than 20°), though steeper dips (60-70°) occur in the Barmouth area. Cleavage strikes N-S and is usually steeply dipping. There are several major faults, the majority of which are orientated N-S or NE-SW. Major N-S faults include the Trawsfynydd Fault, which forms the approximate eastern limit of the exposed Harlech Grits Group. Smaller scale faults also affect the area and in general, appear to radiate from the centre of the Dome so that NW-SE and N-S faults predominate in the southern part of the Dome while E-W faulting is more important in the E (BGS 1:50,000, Sheet 135).

The Cambrian succession was metamorphosed to greenschist grade (Greenly 1897; Woodland 1938a, 1939). The main phase of metamorphism and deformation was probably end-Silurian by analogy with other Lower Palaeozoic rocks in Wales, but some deformation may have been earlier. The Cambrian strata of the Harlech Dome form a conformable sequence, but on the eastern side of the Dome they were

FIG 4.1 Geological map of the Harlech Dome.



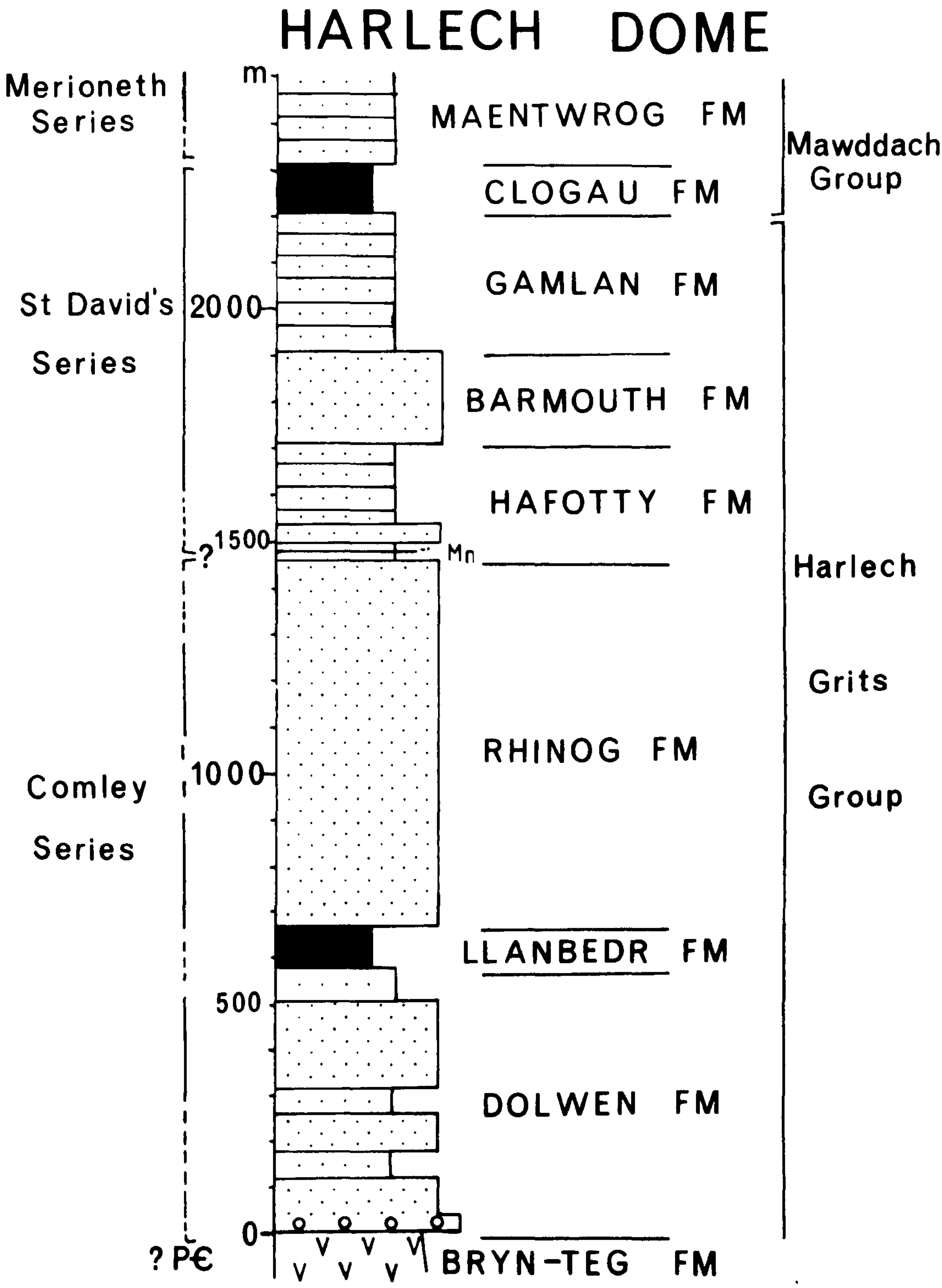
folded (with N-S trending fold axes) and unconformably overlain by the Tremadocian Rhobell Volcanic Group (Wells 1925; Kokelaar 1977, 1979). The Rhobell volcanics are then unconformably overlain by Arenig strata. However on the northern side of the Dome a sequence of Arenig age overlies Tremadocian rocks conformably or disconformably with only a short time gap between the two sequences (Fearnsides 1910b; Fearnsides & Davies 1944). Thus the relationship between the Arenig and older strata is very variable. Localised volcanogenic uplift results in an unconformable relationship around Rhobell Fawr (Kokelaar 1979) which contrasts with the more continuous succession near Porthmadog, north of the Dome.

The Harlech Dome was first described in detail by Sedgwick (1843) who included the area within his "Merioneth Anticlinal" structure. The Geological Survey published maps of the area in 1854, showing the junction between the sandstone dominated "Harlech Grits" below and finer grained beds above ("Lingula Flags"). The rocks were described and details of major faults and mineral veins were given by Ramsay (1866, 1881 p19-27).

Later Lapworth identified a stratiform, manganese ore-bed within the "Harlech Grits" and this allowed a division of the sandstone-rich succession above the manganiferous horizon from that below. A detailed Cambrian succession was described by Andrew (1910) and later refined following more detailed mapping by Matley & Wilson (1946). This succession has since been supplemented by information from the Bryn-teg borehole in the eastern part of the Dome; volcanics were found beneath the lowest exposed Cambrian formation: the Dolwen Formation (Allen & Jackson 1978).

The Geological Survey have recently mapped the eastern part of the Harlech Dome and divided the Cambrian into the Harlech Grits Group (below the Clogau Formation) and the Mawddach Group (the Clogau Formation and above), (Allen *et al.* 1981). The Harlech Grits Group was subdivided into formations (as shown in Fig 4.2) which are largely

FIG 4.2 The Lower and Middle Cambrian succession of the Harlech Dome.



coincident with the divisions of Matley & Wilson (1946). The succession is divided as follows (Allen & Jackson 1978, 1985; BGS 1:50,000, Sheet 135):

a) Bryn-teg Volcanic Formation (over 160m).
Interbedded volcanic sediments, tuffs, tuffites and andesitic lavas.

b) HARLECH GRITS GROUP (c.2000m).

i) Dolwen Formation (575m). Conglomerates and pebbly sandstones at the base of the formation pass upwards into packets of greenish-grey feldspathic sandstone (with occasional pebbly sandstones) and siltstones. Cross-bedding is common in some of the sandstone packages.

ii) Llanbedr Formation (90-180m). Grey and purple siltstones and mudstones with occasional beds of fine sandstone.

iii) Rhinog Formation (425-780m). Predominantly thick bedded sandstones including coarse grained and pebbly units. In general siltstones and mudstones make up a relatively small proportion of the succession. Graded bedding is common.

iv) Hafotty Formation (170-300m). Mainly grey siltstones with occasional sandstone beds. The Manganese ore-bed occurs near the base.

v) Barmouth Formation (60-230m).
Predominantly thick bedded, coarse grained sandstones. Graded bedding is common.

vi) Gamlan Formation (230-360m). Grey, green and purple siltstones, mudstones and very fine sandstones; they are commonly graded and include rare coarse grained, graded sandstones.

c) MAWDDACH GROUP (over 2000m). Dark silty mudstones at the base (Clogau Formation), followed by lighter coloured silty mudstones and siltstones.

The first diagnostic fossils from the lower part of the Cambrian of the Harlech Dome were found in the Clogau Formation (Ramsay 1881). These trilobites indicate a middle Menevian (St David's Series or Middle Cambrian) age (Cowie *et al.* 1972). *Platysolenites* was found in the Dolwen Formation and indicates a probable Lower Cambrian age (Rushton 1978), so the underlying Bryn-teg Volcanic Formation may be Precambrian or Lower Cambrian in age. The Llanbedr Formation contains brachiopods (Lockley & Wilcox 1979) which confirm its age as Lower Cambrian or younger. The manganese ore-bed in the Hafotty Formation is probably late Lower or early Middle Cambrian in age since late Lower Cambrian trilobites occur directly below manganiferous shales in a similar sequence on St Tudwal's Peninsula (Bassett *et al.* 1976).

Previous research in the Harlech Dome has concentrated on particular aspects of the geology. Powell (1955), as a result of a gravity survey, located the Mochras Fault and suggested that Precambrian basement underlies the Harlech Grits Group. The petrography of the Harlech Grits Group have been studied with particular reference to provenance and metamorphism (Greenly 1897; Woodland 1938a,b,c, 1939; Okada 1966, 1967; de Bethune 1972). Mineralisation has also been studied in detail; this includes work on the petrography and geochemistry of the manganese ore-bed (Woodland 1939; 1956; Mohr 1955, 1956a,b, 1959, 1964; Mohr & Allen 1965; Glasby 1974; Binstock 1977; Bennett 1987), copper mineralisation (Rice & Sharp 1976; Allen & Easterbrook 1978), gold mineralisation (e.g. Andrew 1910) and the affects of the Ordovician intrusions on Cambrian sediments (Allen *et al.* 1976). Tectonic structures within the Upper Cambrian have also been examined (Hawkins & Jones 1981; Hawkins 1985) and the geomorphology has been described (Foster 1968), the latter of which included the identification of structurally produced lineaments.

The sedimentology of the Harlech Dome has also received recent attention, as follows.

Bailey (1930) divided arenaceous rocks into two facies:

a current (or cross) bedded facies and a graded bedded facies. The former he interpreted as being deposited in predominantly shallow water while the latter he envisaged as being characteristic of deeper water deposits. Two of the localities cited to illustrate the graded facies were from the Harlech Grits Group at Barmouth and Harlech. Kuenen & Migliorini (1950) identified the link between turbidity currents and graded bedding and thus regarded turbidity currents as the main transporting agents of the graded facies. This link was used by Kuenen (1953) and Kopstein (1954) to infer that the Harlech Grits Group above the Llanbedr Formation were deposited in a geosyncline by turbidity currents. Crimes (1970a) reaffirmed this view, basing his interpretation on the following evidence. The:

i) alternation of greywacke sandstone and shale beds.

ii) constancy in thickness of individual beds.

iii) general absence or scarcity of fossils.

iv) presence of graded bedding.

v) presence of sole marks.

vi) angularity of grains.

vii) generally poor sorting.

Crimes also used Walker's (1967) criteria for determining proximality in turbidites to infer that the Rhinog and Barmouth Formations are dominated by proximal turbidite sandstones.

There has been considerable debate over the source(s) and palaeocurrents of the Harlech Grits Group (reviewed by Stubblefield 1956; Bassett 1963). Greenly (1897, 1919) argued that many of the clasts from the Harlech Grits Group were probably derived from rocks similar to the Mona Complex, as exposed on Anglesey, thus implying a northerly source. This view was supported by Woodland (in Matley & Wilson 1946) on the basis of the petrography of the Harlech Grits Group, though an easterly source was also favoured (Matley & Wilson 1946) based on the thickening of certain sandstone packets. Matley & Wilson suggested that the Harlech Grits were derived from a source similar to the Mona

Complex which lay to the east or northeast of the present outcrops. Jones (1938, 1955) proposed sources to the west, northwest and north of the Harlech Dome, based on regional facies and thickness variations within a deep NE-SW elongated geosyncline.

Kuenen (1953) and Kopstein (1954) argued for a south-south-westerly source for most of the Harlech Grits, based mainly on grain orientations, though some sole structures and cross-lamination were also used. However Voll (pers. comm. in Bassett & Walton 1960) pointed out that grain orientations were aligned parallel to the cleavage and therefore were controlled by end-Silurian deformation rather than by early Cambrian depositional processes. Kuenen and Kopsteins' other palaeocurrent data were regarded as inconclusive.

Knill (1959) proposed a northeast to north-north-easterly source for the Rhinog Formation on the basis of palaeocurrents from sole marks and cross-lamination. These structures indicated flow mainly transverse to the basin margins while the Hell's Mouth Grits to the west (of similar age) flowed axially (Bassett & Walton 1960). Knill also suggested that there was flow from the south during Barmouth Formation times on the basis of palaeocurrents from sole structures. A southerly source had also been postulated by Matley & Wilson (1946), based on variations in thickness of the Barmouth Formation.

Woodland (1939) and Mohr (1955, 1956b, 1959 & 1964) also argued on the basis of petrography and geochemistry that the Hafotty Formation and the manganese ore-bed in particular, were derived from the erosion of source lands to the east and northwest.

Crimes (1970a) produced a more detailed pattern of palaeocurrents for the Cambrian rocks of the Welsh Basin. His results for the Harlech Grits Group are as follows :

i) Rhinog Formation. Sole structures at the base of graded beds are aligned N-S; flutes indicate supply from the north. However ungraded conglomerate and coarse sandstone filled scours are commonly aligned E-W and are often associated with tabular cross-bedding indicating transport predominantly from the east.

ii) Barmouth Formation. NNW-SSE orientated sole marks occur, including flutes which indicate a southerly supply. There is also a minor NW-SE alignment of some scours.

iii) Gamlan Formation. Coarse turbidites flowed mainly towards the north-west, while finer grained turbidites flowed north and north-east.

4.2 : Bryn-teg Volcanic Formation.

The Bryn-teg borehole, situated near the axis of the Dolwen Pericline (Fig 4.3), revealed 160m of Bryn-teg Volcanic Formation beneath the oldest exposed strata of the Harlech Dome (the Dolwen Formation). The Bryn-teg Volcanic Formation is composed of andesitic and basic lavas, volcanic sandstones, tuffites and tuffs as well as coeval basic intrusives. The succession (Fig 4.4) can be divided into several units (Allen & Jackson 1978):

1) Andesite lava flows. The andesites contain albite phenocrysts, are amygdaloidal and locally autobrecciated. The lavas are interbedded with minor amounts of tuffite and volcanic sandstones.

2) Rhythms comprising of:

a) Coarse grained tuffite beds. They show coarse-tail, normal and reverse grading and were probably deposited from volcanic airfall eruptions with some redeposition by mass-flow processes.

b) Interbedded volcanic sandstones and mudstones. These beds are usually ungraded and occasionally contain flaser bedding which indicate that they were probably deposited in shallow water.

c) Siltstones and fine sandstones. These beds are often graded and contain top-absent Bouma sequences (Walker 1965) which indicate relatively proximal turbidite deposition. Thin tuffs are also interbedded which were deposited from fine ash eruptions.

3) Dacite lava.

4) Volcanic sediments. This unit contains both graded and ungraded sandstones which are rich in volcanic clasts and are interbedded with siltstones.

5) Tuffitic siltstones and volcanic sandstones. This unit and unit 6 are intruded by numerous basaltic intrusions which may be coeval with the extrusives.

6) Tuffites. The lower parts of beds are largely ungraded while the upper parts tend to fine and thin upwards. This feature is interpreted by Allen & Jackson as

FIG 4.3 Geological Map of the Dolwen Pericline.

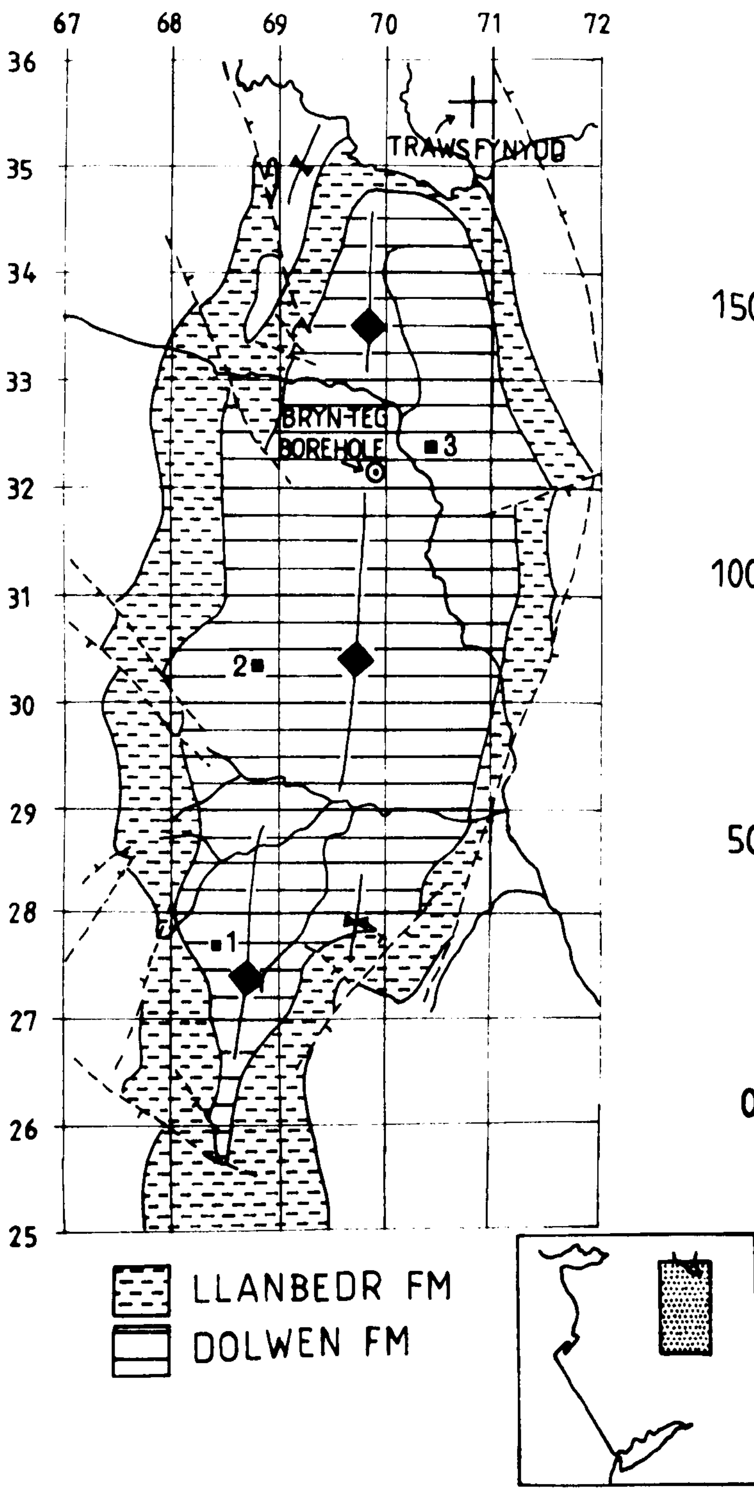
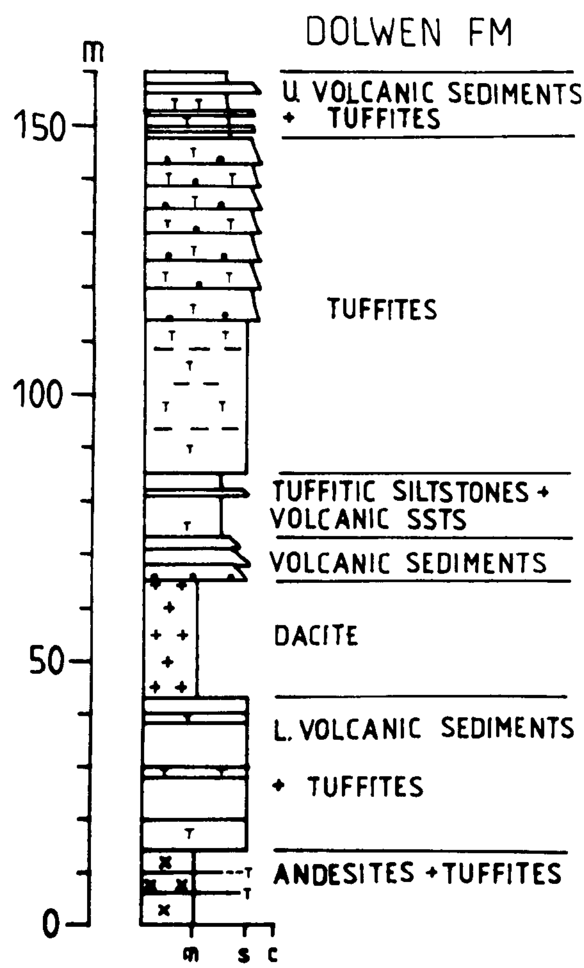


FIG 4.4 Schematic log

of the Bryn-teg Volcanic Fm.



T tuffite

resulting from separation within a turbidity current into a high density part producing "proximal" turbidite deposits and an upper more "distal" part produced by lower density flow (*sensu* Walker 1967).

7) Upper Volcanic sediments and tuffites. This unit contains tuffites and sandstones, siltstones and mudstones, most of which have a volcaniclastic component.

The Bryn-teg Volcanic Formation contains a range of lithologies which include andesitic and basic lavas, pyroclastic deposits and redeposited volcanic sediments. This sequence represents basic and intermediate volcanic activity in a subaqueous environment. Mass-flow processes were probably important on unstable slopes dominated by pyroclastic debris (Allen & Jackson 1978).

4.3 : Dolwen Formation.

The Dolwen Formation is the lowest formation in the Harlech Grits Group. It is exposed in the central part of the Dolwen Pericline, on the eastern side of the Harlech Dome (Fig 4.3). The Dolwen Pericline forms generally low-lying, boggy terrain in which outcrops are few and mainly poor. A small exposure of Dolwen Formation has also been mapped on the western side of the Dome near Llanbedr [SH 583 260] (BGS 1:50,000, Sheet 135).

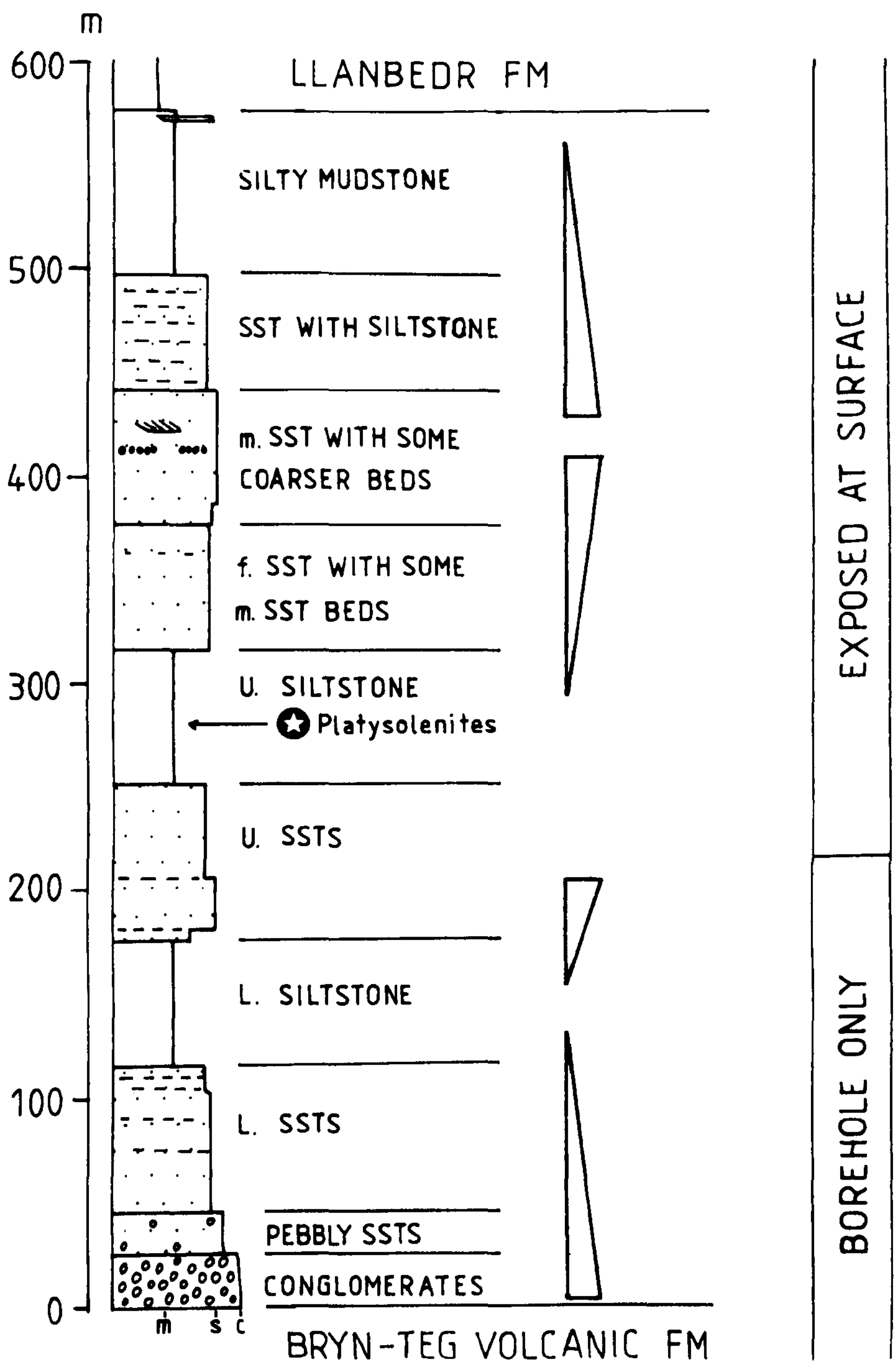
The base of the Dolwen Formation is defined in the Bryn-teg borehole (Allen & Jackson 1978). The Dolwen Formation contains a large scale fining and thinning upward sequence at the base followed by a series of alternating coarsening and fining upward sequences (Fig 4.5). The Dolwen Formation was divided into a series of units by Allen & Jackson (1978). Units 1 to 4 and part of 5 do not outcrop at the surface and are described from the details of the Bryn-teg borehole published by Allen & Jackson. The sequence is as follows:

1) Conglomerates.

The base of the Dolwen Formation is marked by conglomerates which rest on volcanic sandstones and siltstones of the Bryn-teg Volcanic Formation. There is no angular unconformity at the contact. The basal conglomerates of the Dolwen Formation lack clasts of Bryn-teg type but are rich in acidic volcanic clasts, suggesting that a new source region was uplifted and eroded after deposition of the Bryn-teg Volcanic Formation. This abrupt change of source, together with the sharp facies change from turbidite sandstones and siltstones to conglomerates may indicate that the Dolwen Formation is disconformable on Bryn-teg Volcanic Formation.

The clasts in the conglomerate are angular to sub-angular and are poorly sorted. Cobble conglomerates predominate near the base of the unit while pebble conglomerates and pebbly sandstones are more common near the

FIG 4.5 Schematic log of the Dolwen Formation.



top. Bedding is also more clearly defined towards the top of the unit. This unit therefore shows a marked fining and thinning upwards trend.

2) Lower Sandstones.

The fining upwards trend continues in this unit. Thin pebble beds decrease in abundance upwards within the Lower Sandstones and the upper part is in general thinner bedded. Smaller scale (of the order of a few metres) fining upward cycles can also be identified and contain the following divisions:

a) Scour and fill structures, often infilled by coarse sandstone lags containing intraclasts.

b) Large scale cross-bedded sandstones.

c) Finer grained sandstones containing parallel lamination and cross lamination (usually as single sets 2-5cm thick).

d) Mudstones.

Each cycle indicates a period of gradually decreasing flow velocity, possibly associated with the fill of shallow channels.

3) Lower Siltstones.

The lower part of this unit forms the uppermost part of the large scale fining upward sequence as beds generally thin upwards. However near the top of this unit the beds coarsen upwards and pass up gradationally into the Upper Sandstones.

4) Upper Sandstones.

The Upper Sandstones are of similar type to the Lower Sandstones (type 2).

5) Upper Siltstones.

This unit consists of siltstones intercalated with silty mudstone which coarsen upwards into the overlying unit. It also contains *Platysolenites*, a siliceous worm tube (Rushton 1978).

6) Fine Sandstones with some medium sandstone beds.

There is good natural exposure of these beds which were studied at three localities (Fig 4.3): north of Cefn Cam (locality 1 [SH 685 277]), Crawcellt (locality 2 [SH 691 304]) and north of Aber (locality 3 [SH 703 327]). Several types of beds occur interbedded together in the same facies (for instance as logged near Cefn Cam, Fig 4.6):

i) Parallel laminated fine to medium sandstones. Beds average 20cm thick with a range of 5-52cm. Some beds are more strongly laminated than others, though more pronounced lamination is generally more common near the top of beds and may include parallel and cross lamination. Sets of cross lamination are usually 1-2cm thick (with a maximum of 6-10cm) and generally contain low angle foresets. Single set cross lamination is most common. Convolute lamination also occurs locally with an amplitude of about 5cm.

ii) Massive fine to medium sandstones. These beds are occasionally faintly laminated and may include a coarse grained lag near their base. Beds have similar average thicknesses to bed type "i", though the maximum bed thickness is greater (79cm).

iii) Coarse to very coarse sandstone. Beds range from 20-27cm thick. These beds may have erosional bases, are often very irregular in thickness and commonly have sharp tops.

iv) Siltstones and mudstones. These beds are usually thinly bedded (1-8cm). They form only a small part of what is mainly a sandstone dominated sequence.

v) Cross-bedded medium sandstones. Sets of cross-bedding normally range from 7-32cm in thickness (commonly about 15-20cm). Trough cross-bedding is most common, but tabular cross-bedding also occurs locally. Cross-bedding may occur as single sets or more commonly as cosets (e.g. Plate 4/I).

Fig 4.7 shows several sets of cross-bedding viewed

FIG 4.6 Log of part of the Dolwen Formation north of Cefn Cam.

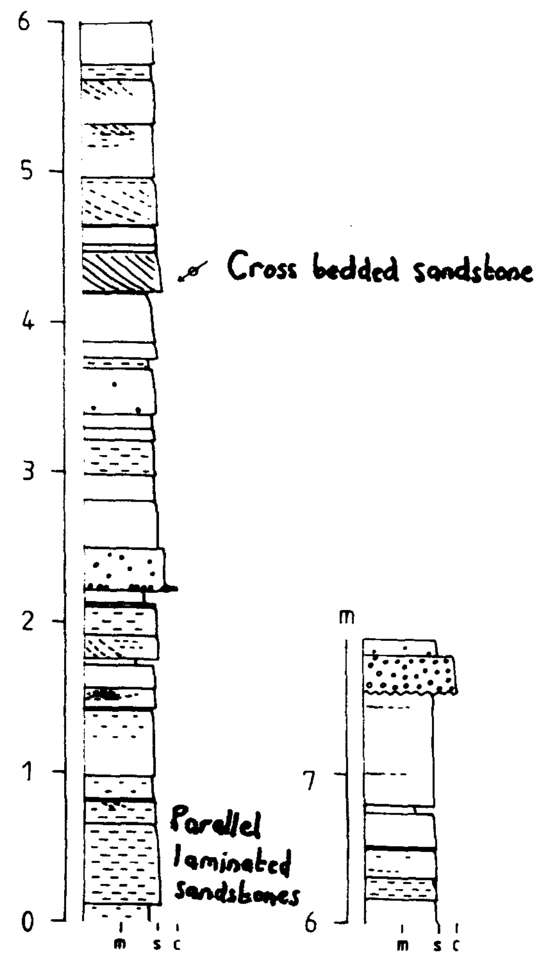


FIG 4.7 Cross-bedding in the Dolwen Formation north of Cefn Cam.

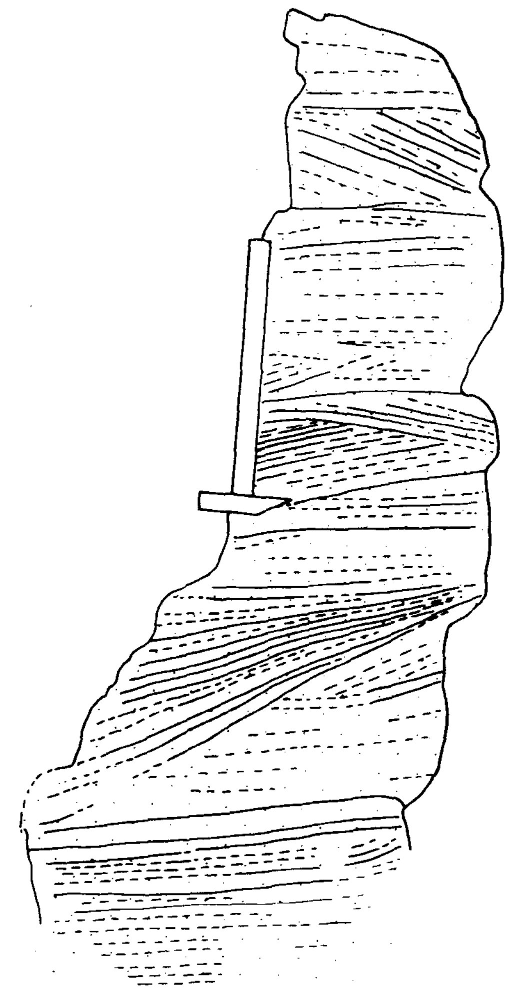


PLATE 4/I : Formset ripple in the Dolwen Formation, north of Cefn Cam.

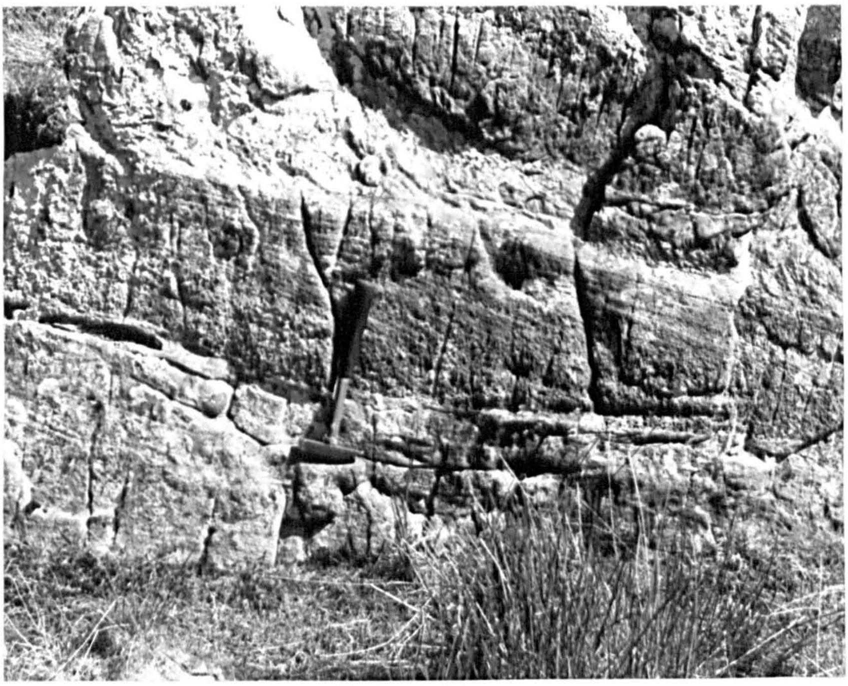


PLATE 4/II : Trough cross-bedding in the Dolwen Formation, north of Aber.

approximately perpendicular to flow. Parallel laminated fine to medium sandstone was eroded and then the scour infilled by 30cm scale cross-bedding. The laminae thin towards the margins of the trough and the slope of the foresets decreases upwards from 30 degrees directly overlying the erosive base to parallel lamination above. The sets of cross-bedding above this are thinner (7-15cm thick) and are probably also trough cross-bedded with parallel lamination between cosets. Thus some cross-bedding occurs within scour fills, some of which were eroded in front of sinuous crested bedforms.

Evidence for the development of particular bedforms may sometimes be preserved. For instance the bedform in Plate 4/I contains low angle cross stratification in its upstream part which mainly aggraded vertically during its initial growth. Steeper foresets then prograded laterally (towards the south-west) with little vertical aggradation in the downstream part of the cross-bedded set. This was followed by more general draping of the structure so that the top of the set is made up of more even, parallel laminae. The bedform is preserved as a formset (35cm amplitude) due to a final blanketing of mud. The bed as a whole has a lensoid geometry thinning rapidly to the south-west and consists largely of parallel laminated sandstone to the north-east. Avalanching of grains resulted in a concentration of coarser grains near the base of foresets.

Fig 4.8 shows a cross-bedded set, the lower set boundary of which erosively cuts into parallel laminated fine sandstone beneath. The coarsest grains (coarse sand and granules) and dark heavy minerals are concentrated near the base of the foresets and along certain laminae. The laminae may have erosive bases and often have a shallow lensoid geometry. Upper and lower boundaries are usually sharp but there may be grading along laminae from granule conglomerate/coarse sandstone to medium sandstone.

Coarser laminae may indicate either slightly higher energy events or an increase in the supply of coarser material. During high energy conditions there was a tendency

FIG 4.8 Cross-bedding from the Dolwen Formation north of Aber.

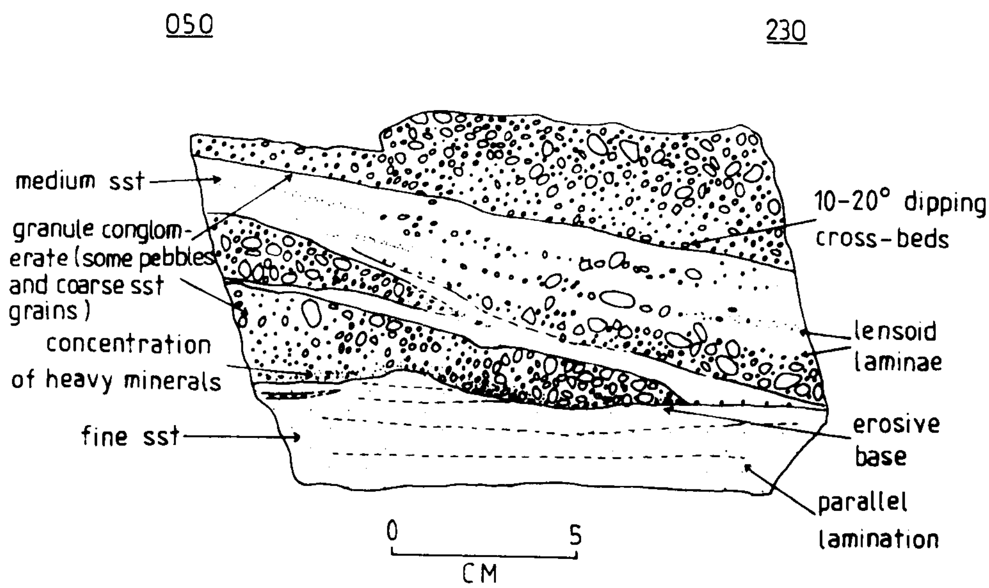
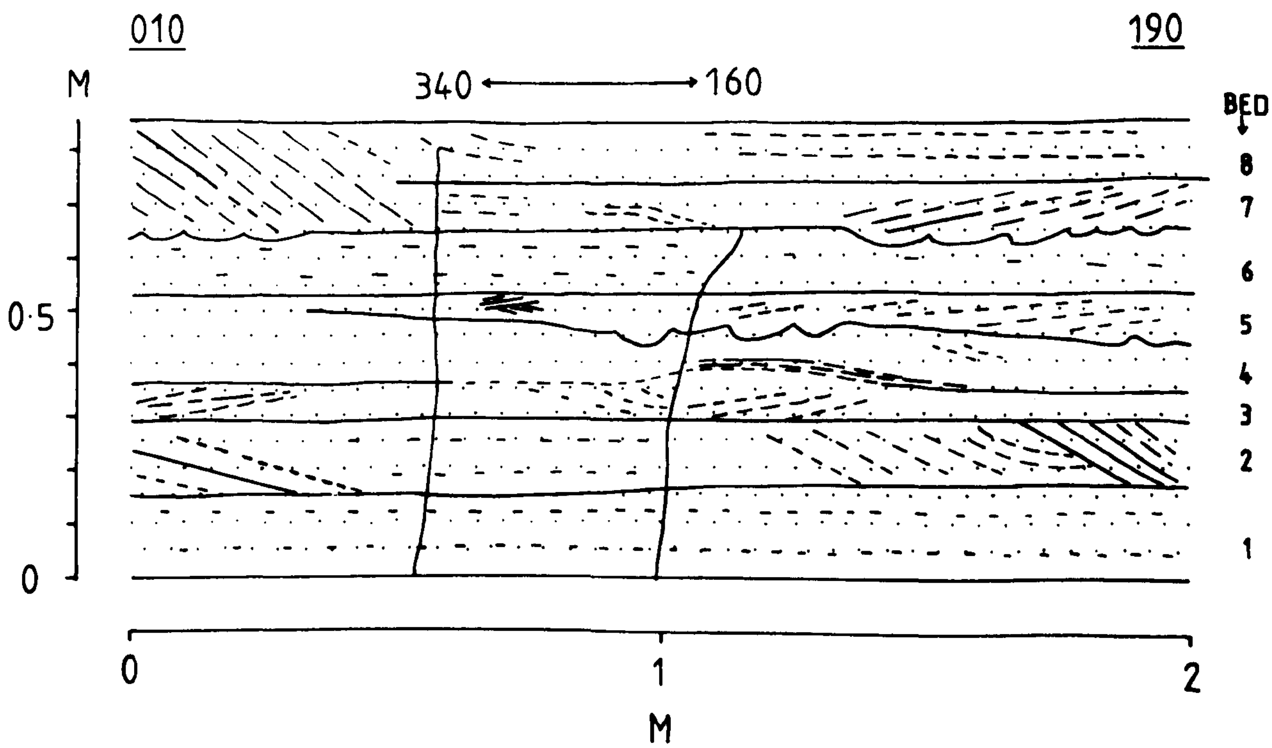


FIG 4.9 Section in the Dolwen Formation north of Cefn Cam.



for foresets to be eroded and then coarser grained sediment to avalanche down the foresets. It is possible that under certain conditions coarse grains may represent coarse lag deposits due to the winnowing effects of currents.

Many beds containing cross-bedding have erosive bases and lensoid geometries on the scale of the outcrops (usually c.20m). For instance bed 5 in Fig 4.9 occurs as a shallow lense with an erosive base and trough cross-bedded sandstone fill. Some beds which are cross stratified pass laterally into parallel lamination. The lateral variability in bed thickness, sedimentary structures and to a lesser extent grain size are shown in a section north of Aber (Fig 4.10) which is aligned oblique to the main palaeoflow direction.

The section in Fig 4.9 also displays highly variable current directions. Some of the cross-bedded sets have steep foresets (up to 34 degrees) dipping mainly to the south-west, though there are a few opposing dips towards the north-east. Cross sets dipping to the north-east are typified by troughs and low angle cross stratification. In the southern part of the section fluctuating current directions are indicated by a trough which is truncated by slightly steeper (30° as opposed to 25°) more tabular foresets, indicating a switch from westward flowing to south-south-west flowing currents. Locally herring-bone cross stratification may be present. Rare three dimensional exposures of cross-bedding suggests flow was predominantly towards the southwest (Fig 4.11).

At Cefn Cam bed types "i", "ii", and "v" are particularly common. Bed types "i" and "v" are common in the section north of Aber, though bed type "iii" becomes more abundant in the upper part of the section.

7) <u>Medium grained sandstone with some coarse beds.</u>

At Cefn Cam another facies is present in addition to that described above which might correspond to unit 7 of the Dolwen Formation. It consists of thinly bedded (on a 1-10cm

FIG 4.11 Dolwen Formation: Palaeocurrents (Cross-bedding).

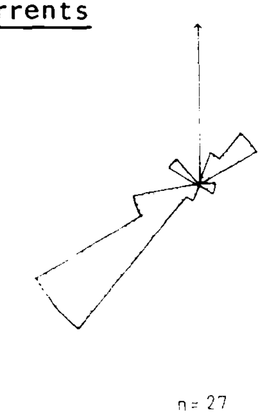
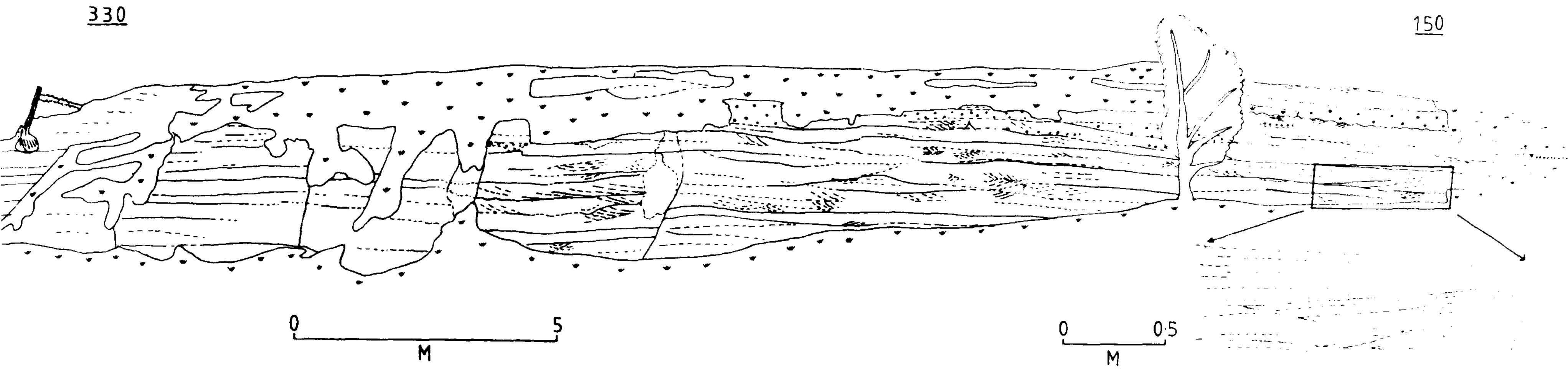


FIG 4.10 Section in the Dolwen Formation north of Aber.



scale) coarse sandstones and granule conglomerates interbedded and/or interlaminated with fine to medium sandstones. Beds normally have abrupt tops and bases though some are graded near the base of beds. The beds are occasionally faintly parallel laminated or cross laminated, often infilling small scours (of the order of 5cm deep). Rare dewatering structures are also present.

This facies may be laterally equivalent to the coarser, upper part of the section north of Aber (where coarse sandstone beds are considerably thicker, c.20-30cm thick) and an isolated exposure of coarse grained, thick bedded sandstone near Crawcellt.

8 and 9) Sandstone with Siltstone and Silty Mudstone.

Near the top of the Dolwen Formation north of Cefn Cam unit 7 type sandstones pass upwards into mainly thin bedded (1-5cm thick) very fine sandstones, siltstones and mudstones, though thicker sandstones occur rarely. Some beds are parallel laminated and possibly show graded tops. Thin graded beds suggest deposition from dilute turbidity currents and a thin ashfall unit is recorded by Allen & Jackson (1985). In general this unit fines upwards and passes up gradationally into the Llanbedr Formation.

Trace Fossils.

Trace fossils occur near the top of the sequence exposed at Cefn Cam. On one bedding plane they form a dense mesh of slightly sinuous horizontal burrows (Plate 4/III). Each burrow is approximately 2mm wide and can be followed for up to 20cm (most for only 5-10cm). They occur on relatively straight crested, symmetrical ripples with a wavelength of approximately 5cm and amplitude of about 1cm. The ripples are composed of fine sandstone, have rounded crests and may be either wave or current produced. Some burrows cross several ripple crests and troughs and therefore post-date the formation of the ripples. A few burrow fills show a thin central divide which may suggest

PLATE 4/III : Trace fossils and ripples, Dolwen Formation, north of Cefn Cam.

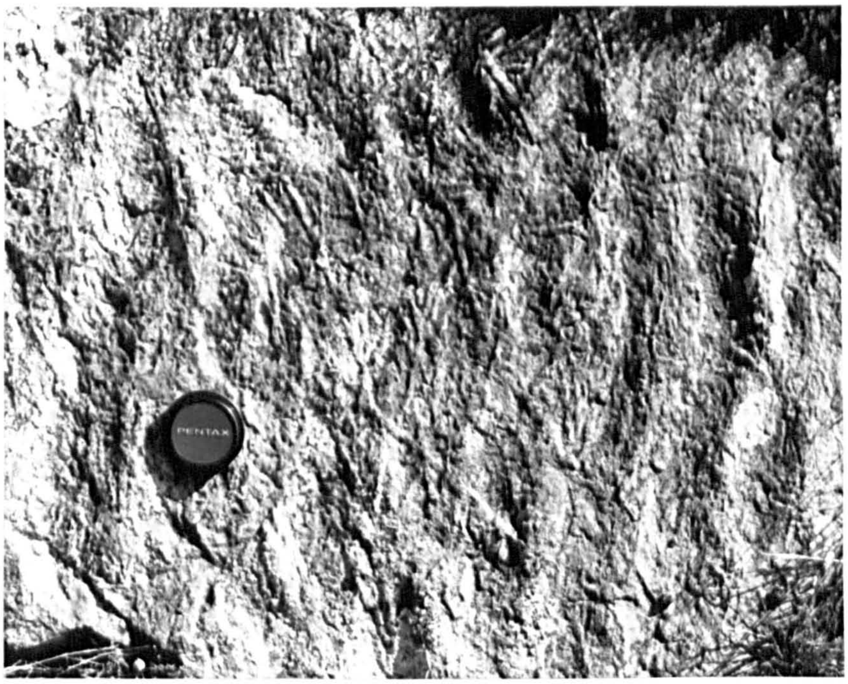


PLATE 4/IV : Radiating burrows, Dolwen Formation, north of Cefn Cam.

that the burrows were originally lined.

Another type was also found (Plate 4/IV). These burrows radiate from a common centre. They are much wider (c.1cm) than the burrows described above and have a maximum observed length of 10cm.

Interpretation.

The lower part of the Dolwen Formation fines upwards from coarse grained, massive conglomerates to sandstones. The smaller scale upward fining cycles of the Lower Sandstones have erosional bases, contain lag deposits and cross stratified sandstones which fine up into siltstones and mudstones. These cycles are similar to upward fining cycles described in alluvial systems with meandering rivers (e.g. Allen 1965). If this interpretation is correct then the gradational passage up from possible mass flow conglomerates to pebbly sandstones, finer sandstones and finally siltstones is likely to reflect a passage from alluvial fan, possibly through braided pebbly sandstone facies to more mature meandering fluvial and siltstone rich alluvial sediments.

The presence of *Platysolenites* in the siltstones overlying the alluvial sediments, might be significant since they may indicate the first marine incursion; *Platysolenites* is predominantly associated with marine sediments (Opik 1956; Hamar 1967; Rushton 1978).

There appears to be a general increase in energy up into the cross stratified facies. This facies is dominated by trough cross sets representing sinuous megaripples. Reactivation surfaces suggest variable flow conditions. Associated are planar beds which are commonly parallel laminated or cross-bedded, often with large changes in grain size between laminae suggesting bedform development in flows of fluctuating flow velocity. Trains of sinuous megaripples form in several fluvial and shallow marine environments but the evidence for bipolar, reversing currents suggests tidal activity (c.f. Allen & Jackson 1978, 1985). The predominance of southwestward directed currents suggest the area was dominated by one set of tidal currents. The apparent absence of channelling at the base of the cross stratified facies suggests tidal shoals rather than deposition in tidal channels.

Above the tidal facies the sequence fines up indicating progressive marine inundation and passes up gradationally into the turbidite dominated Llanbedr Formation.

Therefore the Dolwen Formation indicates a general transgressive deepening trend from possible alluvial fan, fluvial and shallow marine (tidal) deposits up into a turbiditic basinal system.

4.4 : Llanbedr Formation.

The Llanbedr Formation is about 180m thick in its western outcrop, which occurs as a strip of ground elongated north-south, stretching from Barmouth to Harlech. Its other area of outcrop is within the Dolwen Pericline (Fig 4.3) where the Llanbedr Formation is only 90m thick. However in the eastern part of the Dome the formation has been considerably augmented by the intrusion of Ordovician sills, mainly of hornblende diorite porphyry (Matley & Wilson 1946; Rushton 1974). Sandstone beds are more common in the western succession and it is possible that its upper boundary with the Rhinog Formation is diachronous.

The Llanbedr Formation is predominantly composed of purple, blue-grey, dark green and green mudstones, commonly showing a good slaty cleavage. The slates have been worked in small quarries in some parts of the Harlech Dome e.g. Llanfair [SH 580 289], Llanbedr [SH 590 266], Egryn [SH 605 206], Plas Canol [SH 603 191], Moel y Gwartheg [SH 681 319] and Ty Cerrig [SH 684 337]. The cleaved mudstones are interbedded with occasional siltstones and rare sandstones.

At Egryn Quarry much of the bedding is indeterminate but at the eastern end of the quarry a thin bed 10cm thick of parallel laminated green siltstone dips to the east. Higher up in the succession there are occasional sandstones that form a transitional top immediately below the sandstone dominated Rhinog Formation.

At Llanfair, graded sandstones are occasionally interbedded with the slates. The sandstone beds are on average 25cm thick. Small scale clastic dykes are common. Bates (1975) discussed the significance of similar sandstone dykes and injection structures in the Llanbedr Formation at Plas Canol. He argued that the alignment of some of the sandstone dykes parallel to the cleavage indicated that the cleavage was produced during dewatering and initial lithification of soft sediment.

The nature of the upper contact of the Llanbedr Formation varies in different parts of the Harlech Dome. At

Harlech in a road cutting on the A 496 [SH 578 308] the boundary between the Llanbedr Formation and Rhinog Formation is transitional. Two main facies are intercalated:

i) Sandstones, c.30cm thick. Bouma T_{ae} units are associated with very thin interbeds and amalgamated T_a units, though grading is in general poor. Some beds have erosive bases. One bed is 20cm thick, has a highly lensoid geometry and an erosive base. However most beds have a tabular geometry.

ii) Slates. This unit is composed of cleaved siltstones and mudstones and is occasionally parallel laminated (T_d).

This interbedded sequence, which overlies the mudstones in the main part of the formation may represent a coarsening upward sequence associated with the gradual progradation of a sand-rich turbidite system into a mud dominated basinal setting.

However at Ffridd-Llwyd, south of Trawsfynydd, Allen & Jackson (1985) describe greenish grey, cleaved mudstones (Llanbedr Formation) conformably but abruptly overlain by a thick sequence of coarse grained turbidites (Rhinog Formation). Thus there seems to be a much more abrupt change from Llanbedr Formation to Rhinog Formation in the eastern part of the Dome. Initially, sand-rich turbidites may have been deposited earlier in the east relative to the west, possibly in response to tectonic controls at the basin margin. The sand-rich facies then spread diachronously westwards as the turbidite system prograded, thus accounting for the changes in thickness of the Llanbedr Formation and the differences in the nature of its the upper boundary.

The Llanbedr Formation in general therefore represents deposition by relatively dilute turbidity currents with coarser grained, higher density flows becoming more important near the top of the formation.

4.5 : Rhinog Formation, Introduction.

The Rhinog Formation outcrops in much of the central Harlech Dome including a broad area around the Dolwen Pericline which connects with a wide band running N-S from Harlech to Barmouth (Fig 4.1). It is the thickest formation in the Harlech Grits Group and is predominantly made up of sandstones, with few mudstone-rich sequences. Matley & Wilson (1946) calculated that the Rhinog Formation was between 425 and 780m thick. The formation is thickest in a band aligned NE-SW between Barmouth and Trawsfynydd, thinning to the southeast and northwest of this band (Allen & Jackson 1985). However thicknesses are difficult to determine accurately because of variations in the dip of bedding and the affects of faulting.

The Rhinog Formation may be divided into a series of facies:

1) Thin Bedded Facies, mainly composed of relatively fine grained, thin bedded sequences. Base absent Bouma sequences predominate.

2) Conglomeratic Facies, typically composed of thick, metre scale conglomerate beds interbedded with thick bedded, massive and thinner bedded, parallel laminated beds of sandstone. Coarse grained scour fills are common.

3) Amalgamated Coarse Grained Facies, characterised by disorganised sequences of very thick bedded, massive, amalgamated sandstones and coarse scour fills.

4) Sand-rich Facies, predominantly composed of graded and ungraded turbidite sandstones, commonly showing top absent Bouma sequences. Bedding is generally better defined and beds are usually more laterally continuous than the

Amalgamated Coarse Grained Facies. A large variety of bed types are present in this facies including cross-bedded units. It is the most abundant facies in the Rhinog Formation.

Since the Barmouth Formation contains similar facies to the Rhinog Formation, the following discussion will also include examples from the Barmouth Formation.

Facies 1 : Thin Bedded Facies.

Main Characteristics.

The Thin Bedded Facies contains the following characteristic features:

a) Thin beds. Most beds are less than 20cm thick, the majority within the range 1-10cm.

b) Fine grain size. Most sequences of Thin Bedded Facies show alternations of very fine sandstone to coarse siltstone grading into fine siltstones and cleaved mudstones. Thicker beds of fine sandstone may also occur occasionally and coarser sandstones more rarely.

c) A relatively low sandstone-siltstone ratio. This ratio is most commonly below 1 and usually is less than 2, which contrasts with other facies in the Rhinog Formation.

d) Amalgamated beds are generally uncommon. However amalgamation may occasionally occur where sandstones are thicker bedded.

e) Erosive bases of amplitude greater than 2cm are normally absent within this facies. However shallow flutes and other sole structures may be common locally. Thin bedded Facies sometimes fill hollows on the top surface of thick bedded sandstones.

f) Convolute laminated beds may be common.

g) In general base absent Bouma sequences predominate, including: T_{bde}, T_{bcde}, T_{cde} and T_{de}. In general T_a at the base of Bouma sequences is uncommon compared to the other facies in the Rhinog Formation.

h) Trace fossils most commonly occur in this facies.

In general units of Thin Bedded Facies are usually 0-3m thick, intercalated between more thickly bedded sand-rich units. Thicker units of Thin bedded Facies are more common in the generally more poorly exposed, lower parts of the Rhinog Formation and in the western part of the Dome where they form a greater proportion of the sequence. In the western Harlech Dome some siltstone-rich units are sufficiently thick to have been mapped by the BGS (1: 50,000 Sheet 135).

Bed Types.

The Thin Bedded Facies contains the following bed types:

1) Mudstones and siltstones. Often parallel laminated with thin silt laminae (T_{de}).

2) Thin bedded graded beds. These beds have abrupt bases which may be loaded and flamed (Fig 4.12). Beds in general range from 1-10cm thick. Each bed is usually composed of a fine sandstone-siltstone/mudstone couplet. The lower part of the bed (normally composed of very fine sandstone or siltstone) does not usually exceed 1cm thick and this grades up into fine siltstone and mudstone. The sandstone part may be vary in thickness laterally and is usually unlaminated or very faintly laminated. Grading is also evident from cleavage refraction which is often pronounced near the tops of beds.

Many beds do not show development of Bouma sequences above the graded units; instead they occur as graded sandstone-mudstone couplets. However the thicker beds seem to show development of T_d suspension deposits above the sandstone part. This bed type is most directly comparable with beds which occur in Mutti's (1979) D3 subfacies. The lack of lamination and absence of clear Bouma sequences, without significant amounts of traction at the base of the turbidity current suggest relatively rapid deceleration of flow and deposition mainly from suspension.

3) Ripple cross-laminated beds. These beds are composed of very fine sandstone to coarse siltstone, 1-4cm thick. They may be tabular cross laminated (indicating straight crested ripples) or more commonly trough cross

FIG 4.12 Loaded bases, Thin bedded Facies, north-east slopes of Rhinog Fawr.

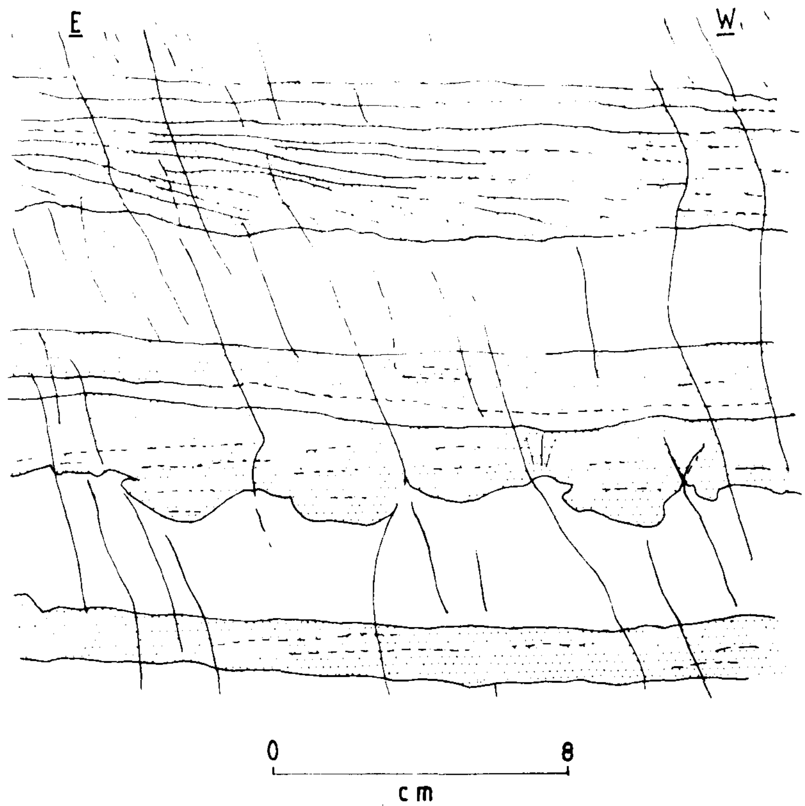
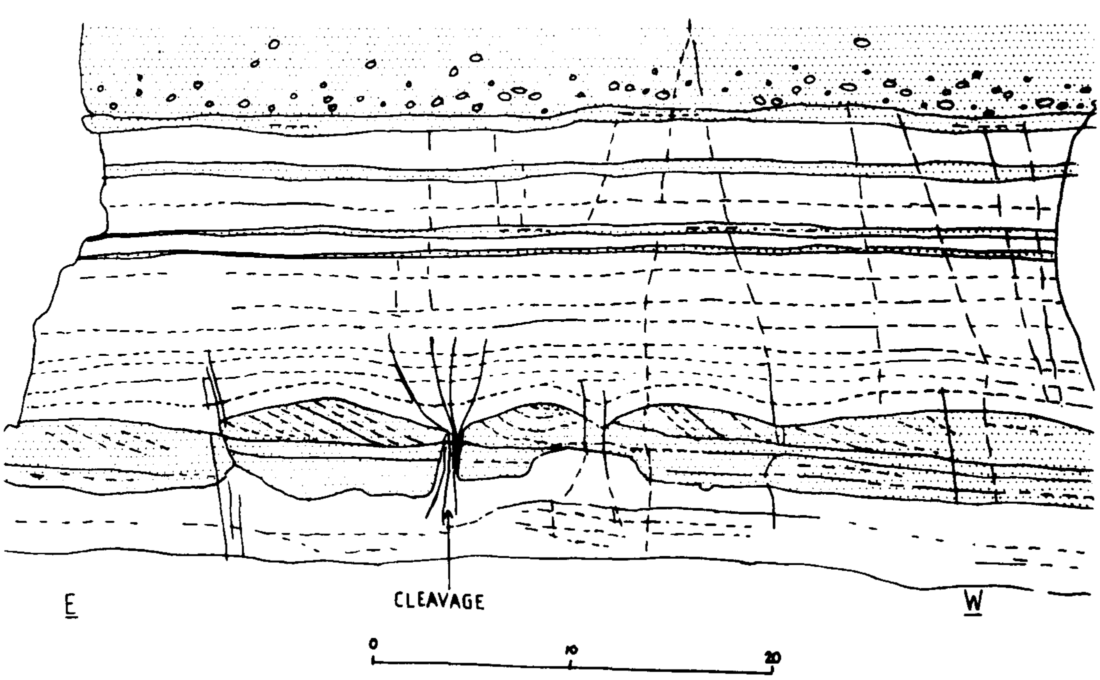


FIG 4.13 Rippled bed, Thin bedded Facies, north-east slopes of Rhinog Fawr.



laminated (indicating sinuous crested ripples), usually as single sets but occasionally as cosets (bed type 4). These beds have abrupt (usually planar) bases and may have graded or abrupt tops. The ripples are often preserved as formsets (Fig 4.13). The ripples have an amplitude of 1-4cm and a wavelength of about 10-20cm. Most ripples are asymmetrical, a few are symmetrical, though the dip of the cross-lamination in a given bed is unidirectional. Palaeocurrents from the dip of ripple cross laminae are variable, though many indicate flow towards the west (Fig 4.14).

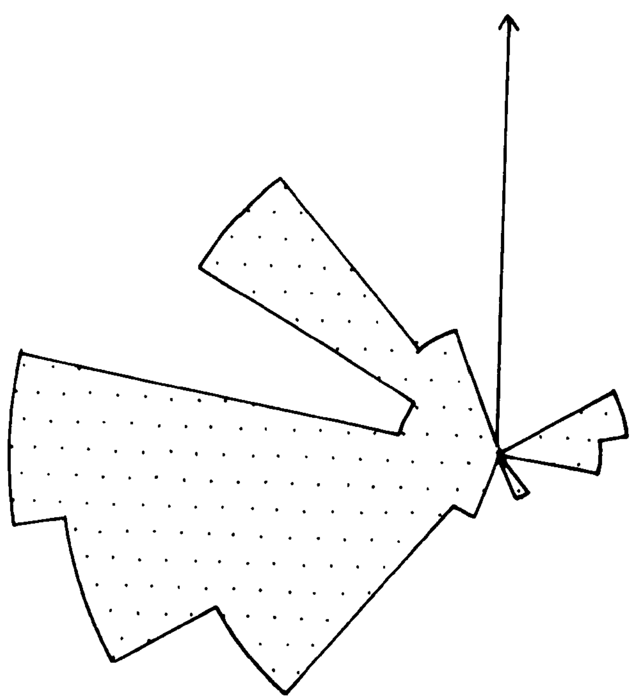
The ripples are commonly overlain by parallel laminated siltstones and mudstones. This division (T_d) is usually less than 1cm thick and passes up into dark green cleaved mudstones. The underlying T_c division often has a sharp top though there may be a thin mud laminae between the T_c and T_d divisions. Therefore Bouma $T_{cd(e)}$ sequences typify this bed type. Some beds contain a thin basal unit of T_b division as part of a Bouma T_{bcde} sequence. However T_b always forms a relatively small proportion of the bed thickness and is more typical of bed type 6.

Some of the ripple cross laminated beds show incipient convolute lamination. These undulatory laminae often have a similar wavelength to the ripple wavelength. Beds containing true convolute lamination are assigned to bed type 5.

When traced laterally the ripple cross laminated beds commonly thicken and thin at a constant wavelength (Plate 4/V). This is probably due to differential compaction of formset ripple trains and may be accentuated by later tectonic-induced boudinage. Locally boudinage has divided particular beds into a series of isolated lenses; cleavage convergence occurs in the inter-lense regions (Fig 4.13). Despite various compactional and tectonic complications these beds are laterally continuous on the outcrop scale (c.10m).

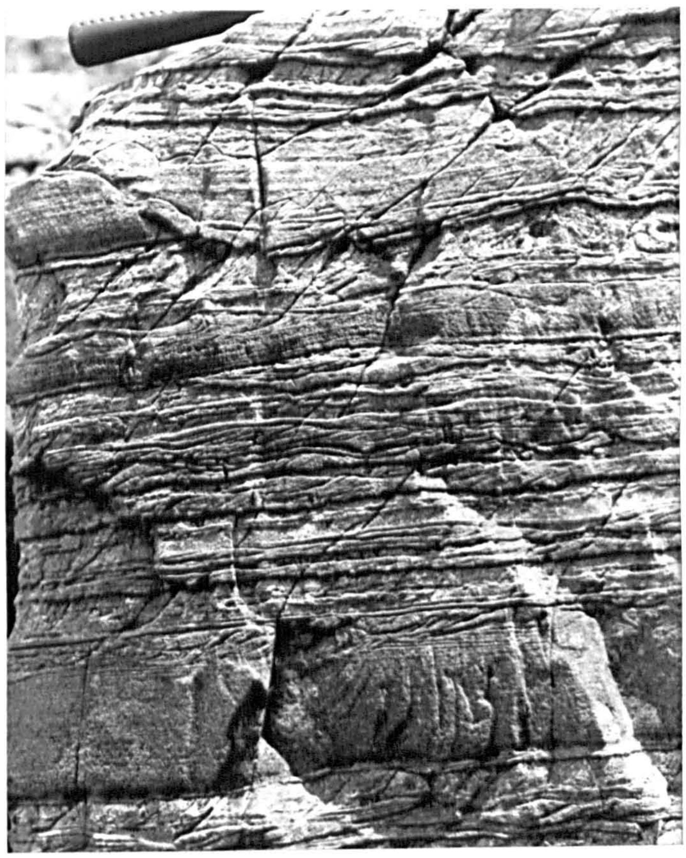
This bed type is therefore represented largely by T_{cde} Bouma sequences and includes only minor amounts of T_{bcde} beds, indicating flow mainly within the lower flow

FIG 4.14 Palaeocurrents from cross-lamination, Thin bedded Facies, Rhinog Formation.



n = 68

PLATE 4/V : Thin bedded Facies, Rhinog Formation, near the Roman Steps..



regime. These beds are similar to bed types which occur in Mutti's (1979) D2 subfacies.

4) Multiple set, cross laminated very fine sandstones. The coarser cross laminated part of these beds ranges from 4cm to about 10cm thick. Bedform amplitudes are of the order of 1-3cm; wavelengths are similar to bed type 3. Double sets are most common, though up to four sets may be present as cosets. Climbing ripples are occasionally present. This bed type includes some T_{cde} and a few T_{bcde} sequences.

5) Convolute laminated beds. These beds are often thicker bedded than many type 3 or 4 beds, though they are commonly derived from Bouma T_c beds, the original set truncations being occasionally preserved. Convolute laminated beds usually range from 6-20cm thick and have narrow, upright anticlines and broad synclines usually affecting the whole of the sandstone part of the sandstone-mudstone couplet. The larger scale folds have a wavelength of approximately 4-10cm which may contain small parasitic folds with a wavelength of 1-2cm. Metadepositional type convolutions (Allen 1977) are most common; convolute laminae merge downwards into parallel lamination, but have abrupt, truncated tops which are overlain by mudstones. Undeformed parallel lamination occasionally overlies convolute lamination; this indicates that the convolute lamination was produced very soon after deposition, i.e. before deposition of the overlying parallel lamination. The deformation was probably produced by dewatering of rapidly deposited sandstones; water escape was concentrated at the narrow convolution crests where laminae thicken. Laminae are occasionally obliterated at the convolution crest, due to the presence of dewatering pipes.

In plan view the convolutions form broad, elongate, N-S aligned troughs with narrow ridges in between, which correspond with the wide synclines and narrow anticlines seen in vertical section (Figs 4.15, 4.16). Whether the N-S elongation was produced as an original soft-sedimentary feature or has been influenced by tectonic deformation is unclear (the ridges run approximately parallel to the cleavage and thus perpendicular to the tectonic shortening). Their N-S elongation is perpendicular to what one would expect from soft-sediment deformation associated with southerly-flowing turbidity currents, though it is possible that this deformation was associated with westerly flowing traction currents (see section on cross-bedding in the Rhinog Formation). There is a tendency for many of the convolutions to overturn with axes dipping towards the west, south and southeast (Fig 4.17 4.18). Since this is the probable direction of flow of most of the currents in the Rhinog Formation (see later), a waning turbidity current may have deformed the convolute lamination.

Other forms of soft sedimentary deformation may also be associated with this bed type:

a) Chaotic lamination. North of the Roman Steps [SH 6638 3047] convolute lamination passes laterally into chaotic lamination. Chaotic lamination may result from the fluidisation of convolute lamination.

b) Sedimentary intrusions. West of F Rock [SH 6617 3143] convolutions, possibly overturned by current drag are intruded by irregular, medium sand filled dykes (Fig 4.19).

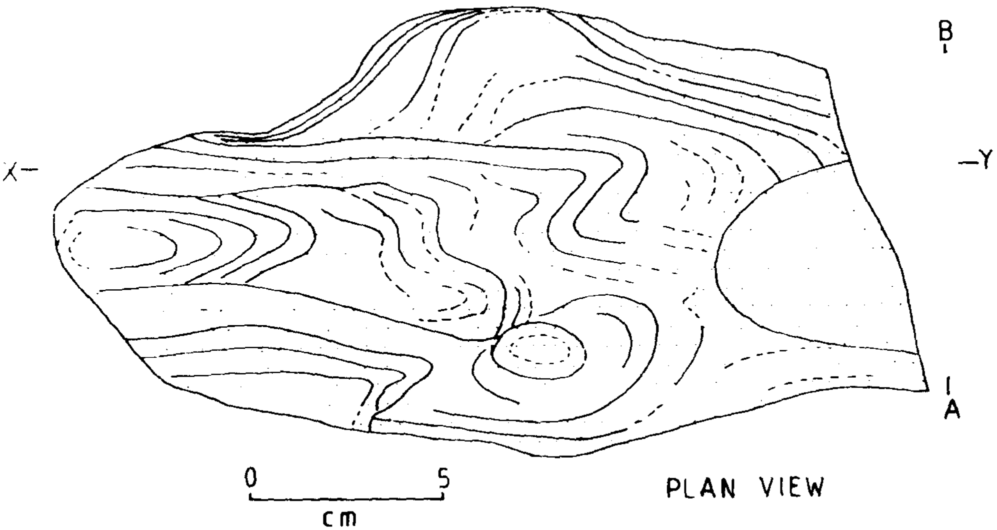
c) Ball and pillow structures, as a result of loading. e.g. west of F Rock [SH 6610 3160] (Fig 4.20).

Bed type 5 probably represents the soft-sediment deformed equivalent of T_{cde} (with some T_{bcde}) sequences. It contains beds comparable with bed types from Mutti's D1 and D2 subfacies.

6) <u>Beds with parallel laminated (T_b) bases.</u>

These beds are predominantly composed of fine to very fine sandstones, have a mean bed thickness of 5-10cm and a maximum bed thickness of 20-30cm. Parallel lamination (T_b) is the dominant Bouma division though this is

FIG 4.15 and 4.16 Convolute lamination, Thin Bedded Facies, north of the Roman Steps.



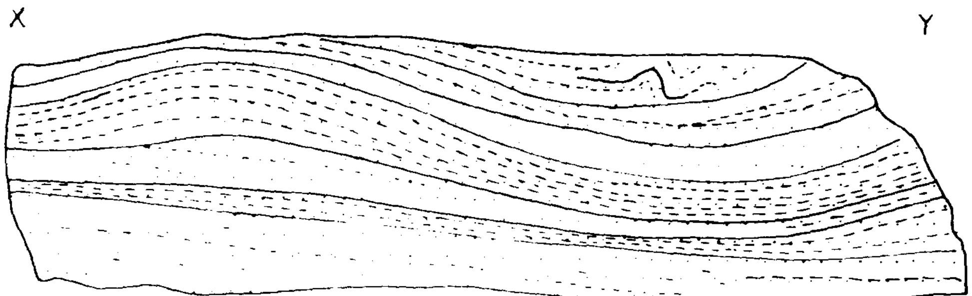
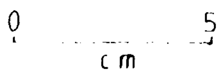
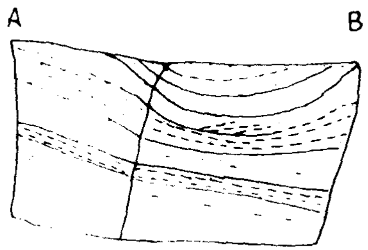


FIG 4.15





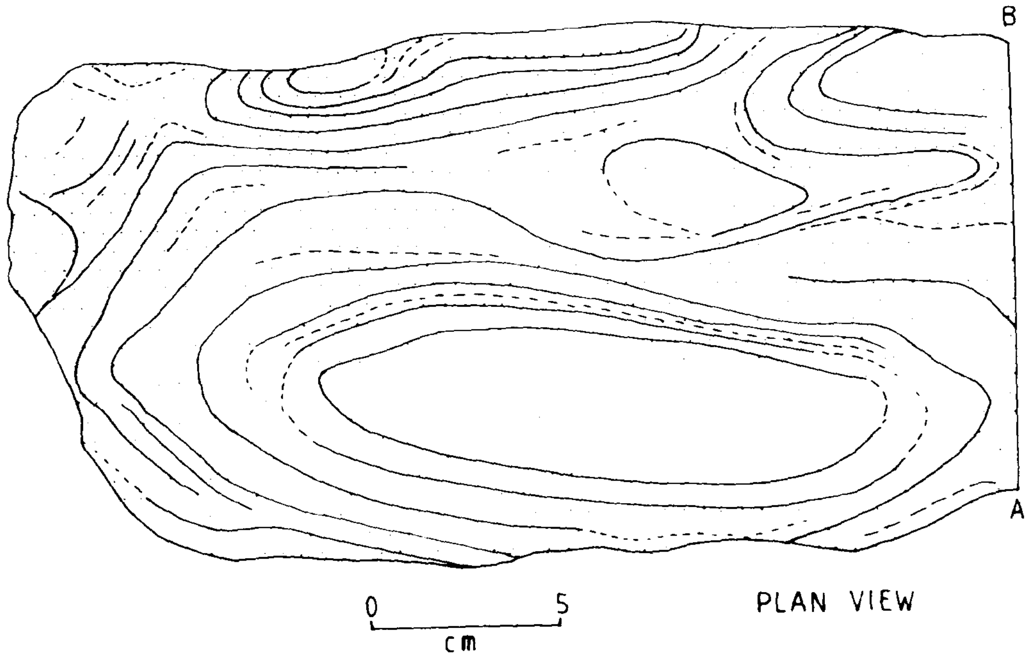


FIG 4.16

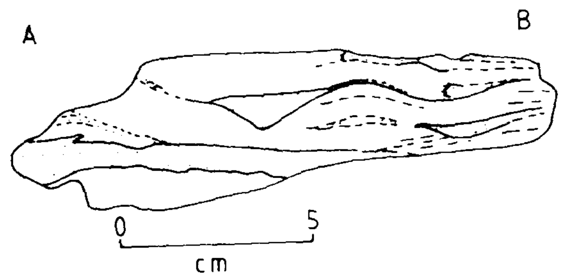


FIG 4.17 Convolute laminated bed, Thin bedded Facies, north-east slopes of Rhinog Fawr.

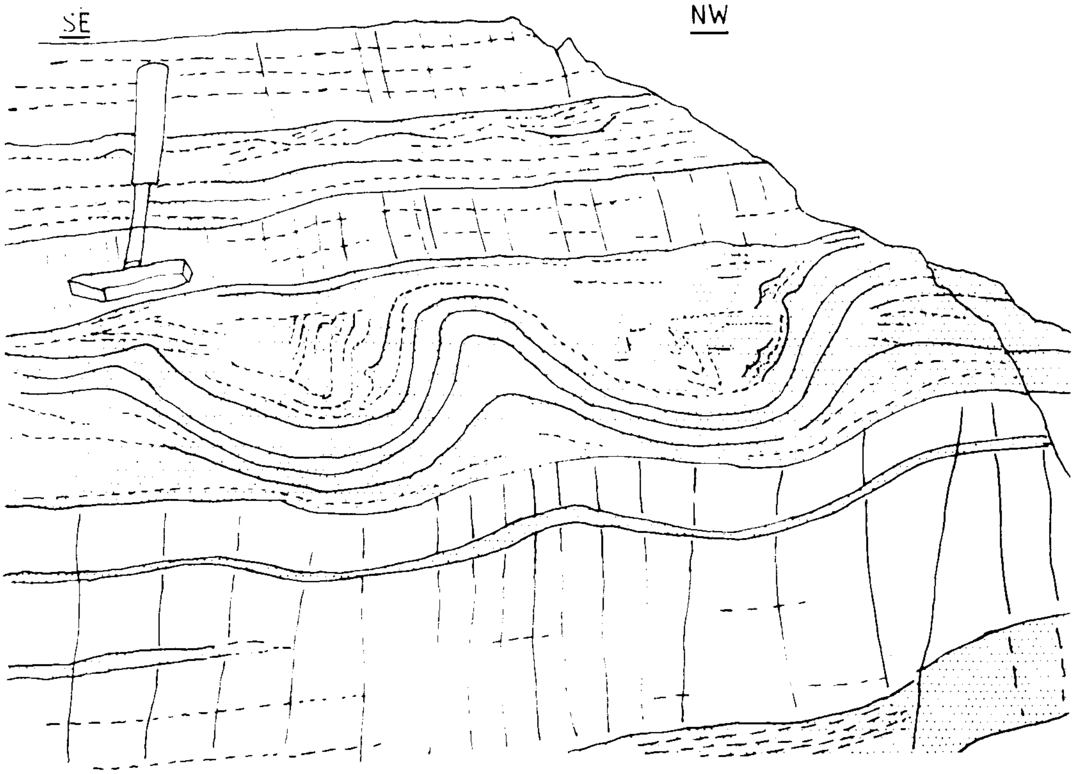


FIG 4.18 Asymmetric convolute lamination, Thin bedded Facies, north-east slopes of Rhinog Fawr.

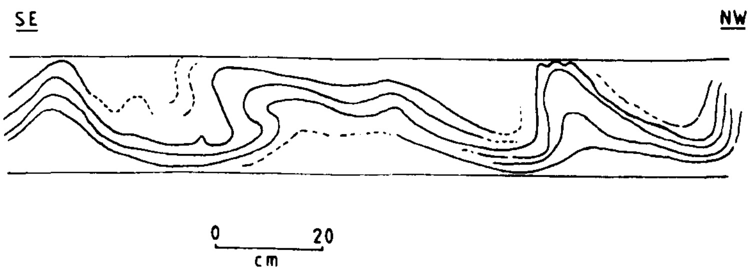


FIG 4.19 Convolutions and sandstone dyke, Thin bedded Facies, west of F Rock.

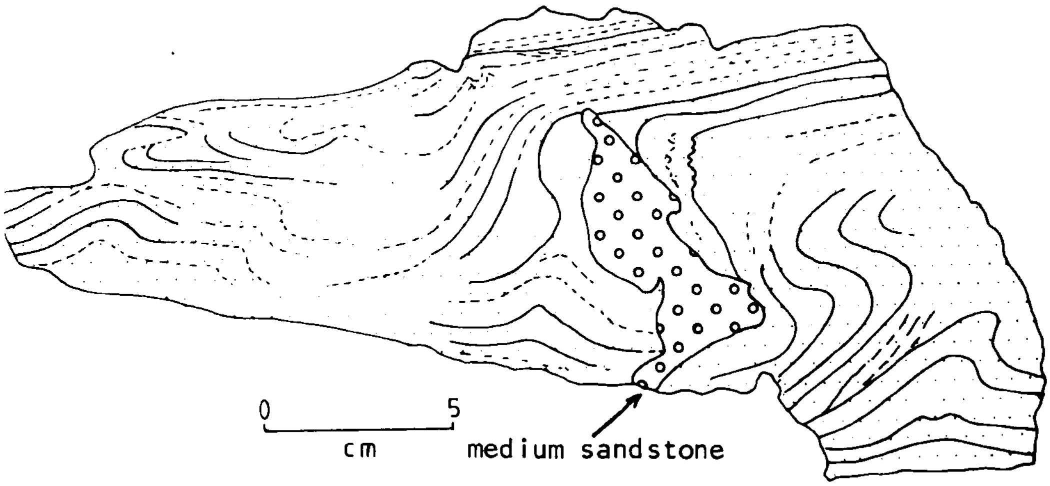
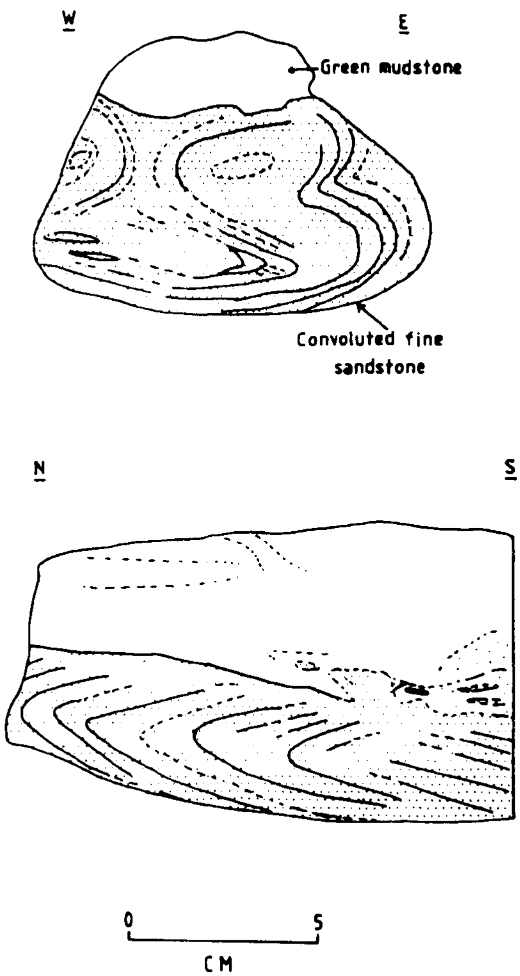


FIG 4.20 Ball and Pillow structures, Thin bedded Facies, west of F Rock.



overlain by cross lamination or convolute lamination (T_c) and laminated siltstones (T_d) (Figs 4.21, 4.22). Base-absent T_{bcde} Bouma sequences are most common, though more rarely T_{bd} middle-absent sequences also occur. One unusual example was found where a 1cm thick cross laminated set occurs below parallel laminated T_b sandstone. This indicates either a slight increase in flow power with time, resulting from an unsteady surging flow during sedimentation from one turbidity current or deposition from two closely spaced turbidity currents resulting in the amalgamation of T_c and T_b divisions.

7) <u>Beds with graded bases (T_a).</u> This bed type is similar to type 6 except that it includes a graded lower part (T_a). Type 7 beds, however are usually composed of medium to fine sandstone and form T_{abcde} Bouma sequences of the order of 20cm thick. This bed type is generally uncommon in most sequences and is similar to the thinner, graded beds of the Sand-rich Facies. However in the Thin-bedded Facies T_a usually forms only a small proportion of a given bed. The graded unit may occasionally be faintly laminated. In some cases the T_b division may be missing, indicative of a middle-absent $T_{acd(e)}$ sequence. This bed type is similar to beds from Mutti's C2 subfacies.

8) <u>Graded beds.</u> Rare T_a beds, usually greater than 20cm thick may be interbedded with the thinner bedded units described above. They typically show top or middle-absent Bouma sequences (T_{ae}), comparable with beds from Mutti's facies C1. These beds are very similar to bed types within the Sand-rich Facies. There are therefore some bed types in common between the Thin Bedded and Sand-rich Facies.

9) <u>Cross-bedded and coarse grained units.</u> This bed type is relatively rare. These beds are usually composed

FIG 4.21 A Bouma Tbcde sequence, Thin bedded Facies, west of F Rock.

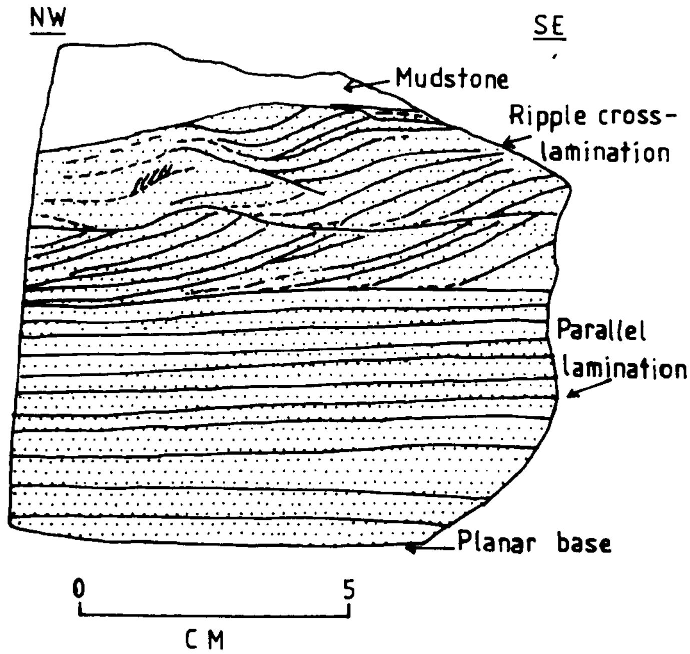
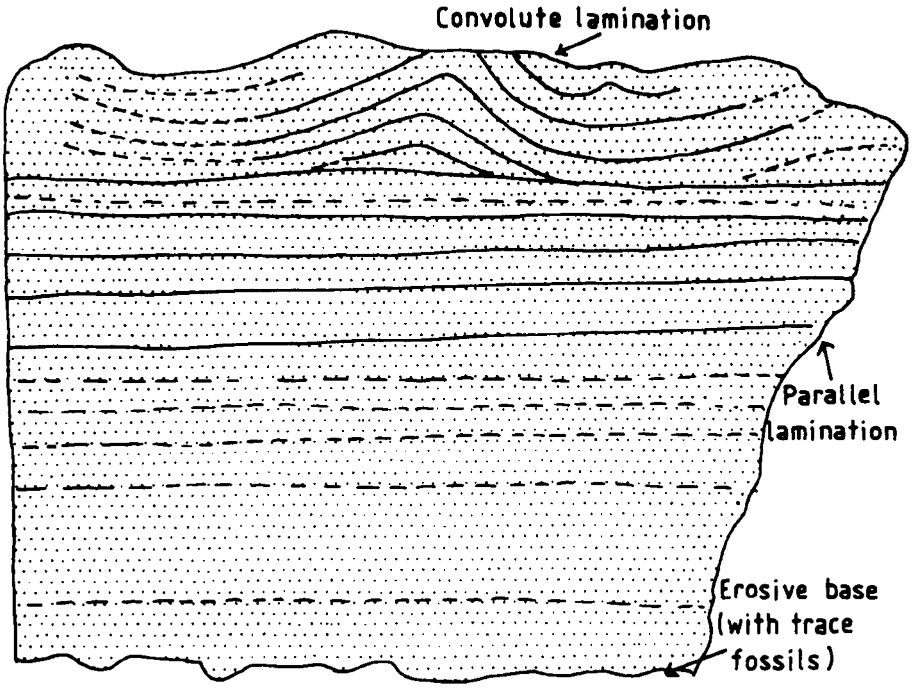


FIG 4.22 A Bouma Tbc sequence with convolute lamination, Thin bedded Facies, west of F Rock.

(scale as above)



of relatively well sorted, coarse or very coarse sandstone, occurring in beds up to 40cm thick. The thinner beds may be cross-bedded throughout (single set), while the thicker beds often have a cross-bedded top. This bed type is often variable in thickness laterally and may have a loaded and flamed base which may also be erosive. For instance one cross-bedded set, outcropping on the northeast slopes of Rhinog Fawr is 8-22cm thick, while another bed nearby has a set height of 6cm. Both sets of cross-bedding indicate flow towards the west. They occur as single, isolated sets, have abrupt upper and lower contacts and do not fit easily into the Bouma model. However these cross beds may be comparable with bed types that occur in Mutti's facies B2 and possibly E.

This cross-bedding does not occur within a modified Bouma sequence (c.f. Allen 1970) and thus was not produced by simple waning flow during deposition of coarse sediment. It is possible that the cross-bedding was produced by tractional flows at the base of dilute, predominantly non-depositing turbidity currents or some other type of traction current. This bed type was deposited from higher velocity currents than the rest of the Thin Bedded Facies. Since the cross-bedded sets have sharp bases and are frequently underlain by siltstones/mudstones the coarse sand was not locally derived. Therefore these flows were sufficiently competent to erode and transport coarse sediment and did not simply winnow previously deposited sediment. Cross-bedding also occurs in the Sand-rich Facies and may result from similar processes (see later).

Vertical Sequences.

The D1, D2 and D3 subfacies of Mutti (1979) may be recognised in the Thin bedded Facies, where certain bed types are clustered in parts of the sequence. However the different bed types are more commonly interbedded in an irregular way (e.g. Plate 4/V). The following generalisations can however be made in order to define subfacies:

D1 This subfacies contains mainly thicker bedded units. The sandstone-siltstone ratio is relatively high and some beds may be amalgamated. Beds with parallel laminated (T_b) bases dominant; bed types 3-9 are present and types 5-7 are particularly common.

D2 This subfacies is more thinly bedded than D1, has a lower sandstone-siltstone ratio and amalgamation is rarer. Sequences are dominated by 3-4cm thick beds of T_{cd}, and occasionally T_{bc}. Bed types 3-8 are present and types 3-5 are common.

D3 This subfacies contains very thin beds of very fine sandstone or siltstone. Parallel laminated beds ($T_{b/d}$) predominate. The sandstone-siltstone ratio is generally very low. Bed types 1 and 2 are common and type 3 is occasionally present.

In general sequences of Thin Bedded Facies show few consistent vertical trends In Plate 4/V, for example consistent trends are lacking; the following Bouma sequences are present: T_{bde}, T_{bcde}, T_{cde} as well as very thin beds/ laminae of $T_{bde/de}$. In log 1 [SH 6618 2970] (Fig 4.23) there is an alternation of subfacies D1 and D2 and the Thin Bedded Facies is erosively overlain by Sand-rich Facies. This sequence also shows few clear vertical trends. Log 2 [SH 6658 3109] (Fig 4.24), however, shows a clear fining and thinning upward sequence approximately 2m thick. The Thin Bedded Facies here shows a transition upwards from subfacies D1 to D2. There is a gradational lower contact with the Sand-rich Facies but a very sharp upper contact. Pronounced fining and thinning upward sequences, similar to log 2 are however, rare in the Rhinog Formation. In general the Sand-rich and Thin Bedded Facies are usually sharply distinct (e.g. log 1).

Soft sediment deformation.

In some places the Thin Bedded Facies is folded. These folds are usually isoclinal, have recumbent fold axes and

FIG 4.23 Log 1, Thin bedded Facies, north-east slopes of Rhinog Fawr.

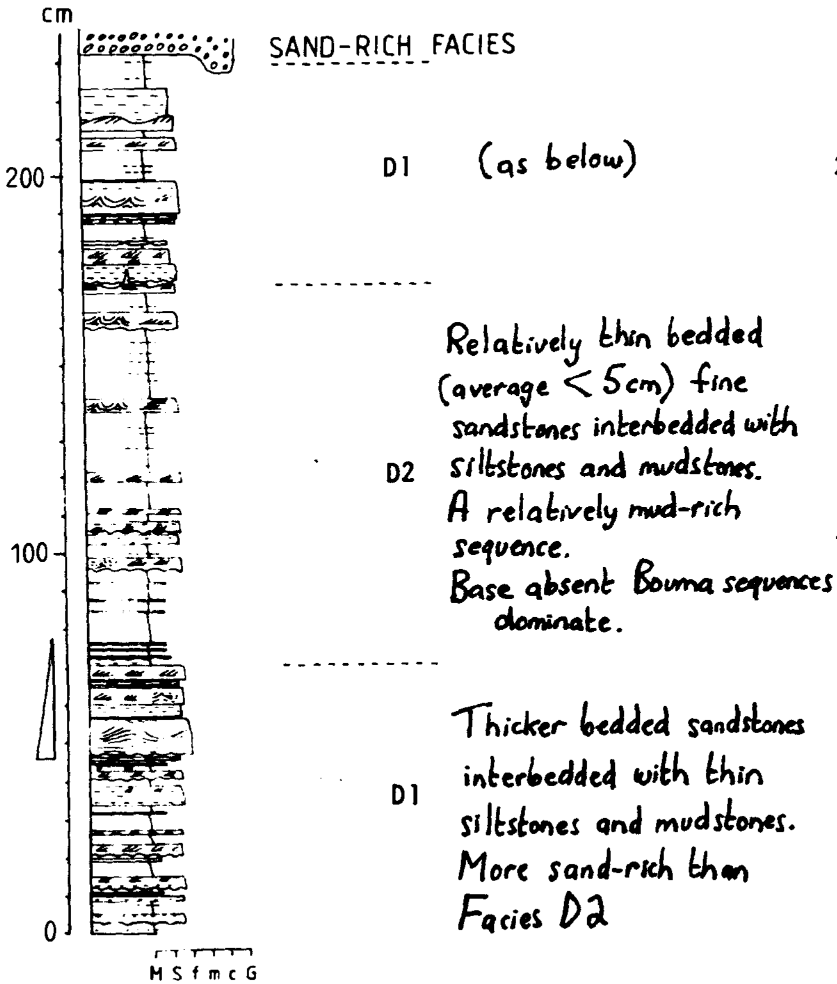
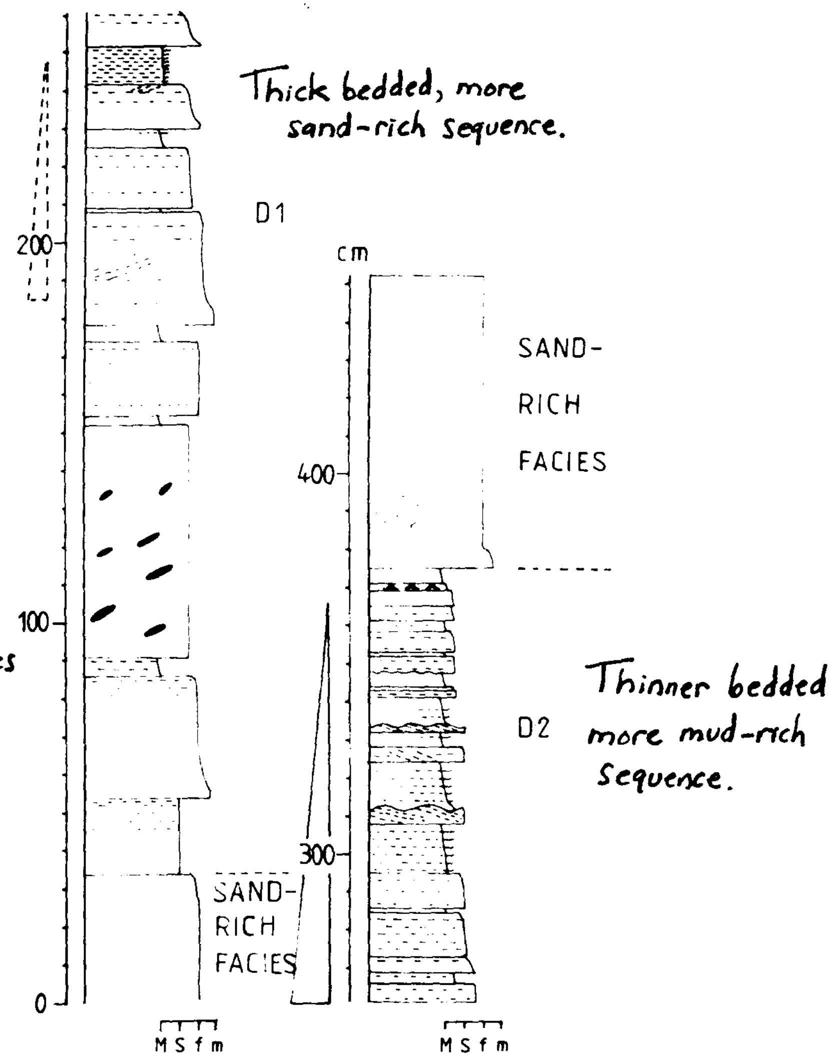


FIG 4.24 Log 2, Thin bedded Facies, F Rock.



may occur within a thin bedded "disturbed" matrix or within a matrix of mudstone or muddy sandstone. At 45m on the Rhinog Fawr section a unit of type 1 beds is folded (Fig 4.25). Since folded laminae are truncated by succeeding beds the folding is therefore of soft sedimentary origin. Similarly at a locality south of the Roman Steps [SH 6556 2990] a 40cm deep, north-south elongated, scour occurs in the upper boundary of a coarse to medium sandstone bed. The scour is unusual in that it is filled with Thin Bedded Facies of D2 type. The Thin Bedded Facies at this locality onlaps over the top of the underlying sandstone. Part of this scour and fill sequence is folded; one recumbent fold in particular (Fig 4.26) indicates either gravitational slumping off the west side of the scour (not seen) or current drag due to flow towards the west. The folded beds are overlain by undisturbed Thin Bedded Facies of D2 type. The scour may have been eroded by a non-depositing turbidity current (i.e. one which deposited its load, possibly of sand-rich type, elsewhere within the turbidite depositional system).

Clasts of Thin Bedded Facies are particularly common within some of the intraclast-rich beds of the Sand-rich Facies which directly overlie the Thin Bedded Facies. Many of the intraclasts show similar structures to those seen in the soft-sediment deformed Thin bedded Facies. It is difficult in some cases to decide whether the thin bedded areas were deformed in situ or were deformed during transport as intraclasts. It is likely that clasts of Thin Bedded Facies could be deformed and transported by high density turbidity currents and deposited as intraclasts within Sand-rich Facies deposits. The intraclasts commonly preserve their original bedding fabric and the long axes of the clasts are usually aligned parallel to this. Therefore evidence of folding within intraclasts might be preserved and folds tightened as a result of later compaction. These processes are probably important in the deformation, erosion and transportation of intraclasts in turbidity currents.

FIG 4.25 Folded Thin bedded Facies, north-east slopes of Rhinog Fawr.

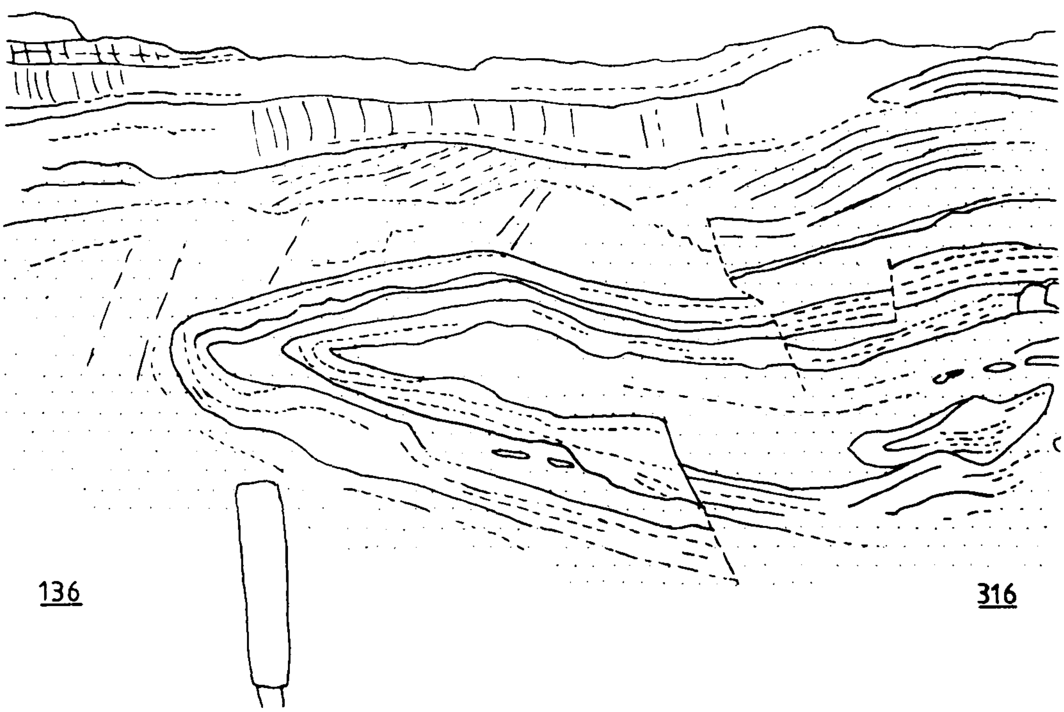
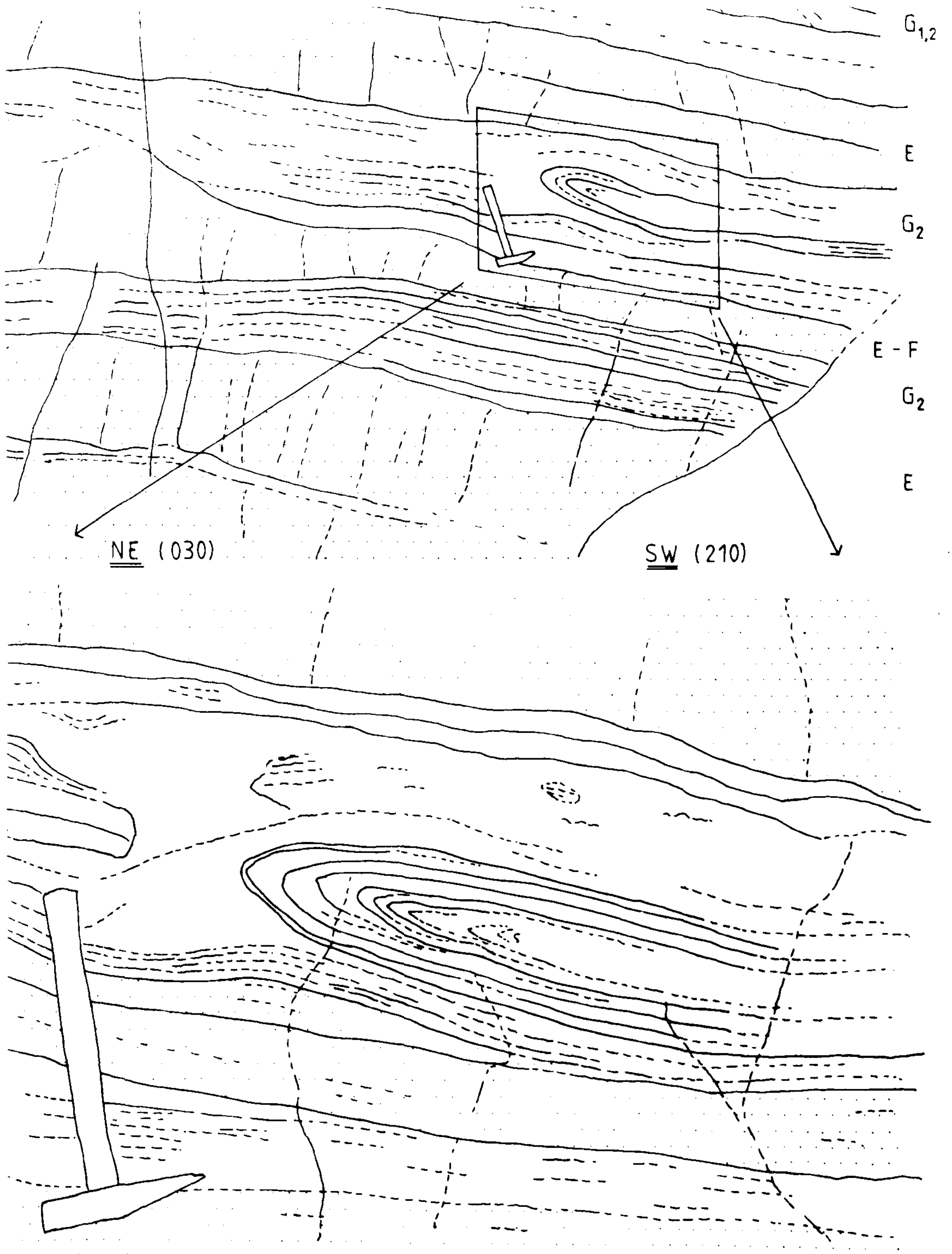


FIG 4.26 Folded Thin bedded Facies, south of the Roman Steps.



Process Interpretation.

The Thin bedded Facies contains predominantly base-absent and middle-absent Bouma sequences which are indicative of waning flow conditions and deposition from relatively dilute turbidity currents. The most important beds types are 2, 4 and 7 which indicate a dominance of T_{bd}, T_{bcd} and T_{cd} Bouma sequences. Erosional structures indicate N-S palaeoflow while cross lamination indicates mainly flow towards the west (Fig 4.14). The main grain size changes occur in the transition between the traction dominated T_c and suspension dominated T_d divisions either by grading or abruptly across a bedding plane. The thicker bedded, coarser grained beds with T_a basal divisions were probably deposited from higher density, higher energy flows.

Facies 2 : Conglomeratic Facies.

Main Characteristics.

The Conglomeratic Facies is the least common of the four facies of the Rhinog Formation. It occurs in the eastern part of the Harlech Dome near Cefn Cam (east of Llyn y Ffran [SH 702 259] and north of Llyn y Ffran [SH 699 260]).

The Conglomeratic Facies is characterised by an abundance of thick, coarse grained beds (including granule and pebble conglomerates), interbedded with medium sandstones. This facies may be transitional with the Amalgamated Coarse Grained Facies. Conglomerate is relatively rare in the other facies, occurring usually as thin scour fills or lenses and forming only a relatively small proportion of most sequences.

Three main bed types can be be identified in this facies:

1) Conglomerates: Granules and pebbles predominate within clast-, or more rarely matrix-supported conglomerates. The matrix is usually comprised of coarse to medium grained sand. Large pebbles are usually the largest grains though most grains occur within the range of granules to medium pebbles. The conglomerates are very variable in thickness, ranging from 2-3cm thick lenses to more laterally continuous beds of the order of 2m thick.

Internal sedimentary structures are rare; little or no clear sedimentary imbrication occurs, though tectonic induced rotation and deformation of grains is sometimes present. Discontinuous flat laminae occasionally occur near the tops of beds and faint parallel lamination can sometimes be detected.

Many of the conglomerate beds infill wide, shallow scours or occur as diffuse, irregular patches. The conglomerate-fills commonly have abrupt tops though some are poorly graded. The scours are often irregular and estimation of palaeoflow is difficult. However the scour in Fig 4.27 was produced by N-S orientated flow. There was either substantial irregular relief within scours or there was also E-W palaeoflow since on N-S oriented faces conglomerate bases also show considerable relief (Fig 4.28). One bed at 6m on Log 1, Llyn y Ffran also has a sharp and highly undulatory top with relief of up to 30cm (Fig 4.29). This feature was probably produced by localised scour of the conglomerate which was subsequently infilled by sand.

2) Thick bedded medium grained sandstones: Beds range from 50cm to several metres thick and are similar to beds from the underlying Sand-rich Facies. Many of the thicker beds are unlaminated or poorly laminated, some show wispy lamination which may have been influenced by water escape as a result of particularly rapid sedimentation. Near the top of the sequence at Llyn y Ffran the sandstones contain pebbly lenses and are transitional in type with the Amalgamated Coarse Grained Facies.

3) Thinner bedded medium grained sandstones: Beds are on average 30cm thick e.g at 5.5m on Log 2, Llyn y Ffran (Fig 4.29). This type is characterised by parallel laminated sandstones (T_b). Cross lamination also occurs locally on the lee side of low amplitude (2-3cm) scours and some of these indicate flow towards the west. Much of the cross-bedding is low angle and tentatively indicates flow towards the west or south-west. The Conglomeritic Facies overlies thick bedded, laterally continuous sandstones of the Sand-rich Facies at Llyn y Ffran and is overlain by laterally variable beds of Sand-rich Facies containing abundant cross-bedding.

Lateral Variation.

Individual beds show considerable lateral variability in thickness and more rarely in grain size. This contrasts with the underlying Sand-rich Facies at Llyn y Ffran which is laterally continuous on the scale of the outcrop (Fig 4.29). However there is no consistent thickening or thinning

FIG 4.27 Conglomerate filled scour, Conglomeratic Facies, Llyn y Ffran.

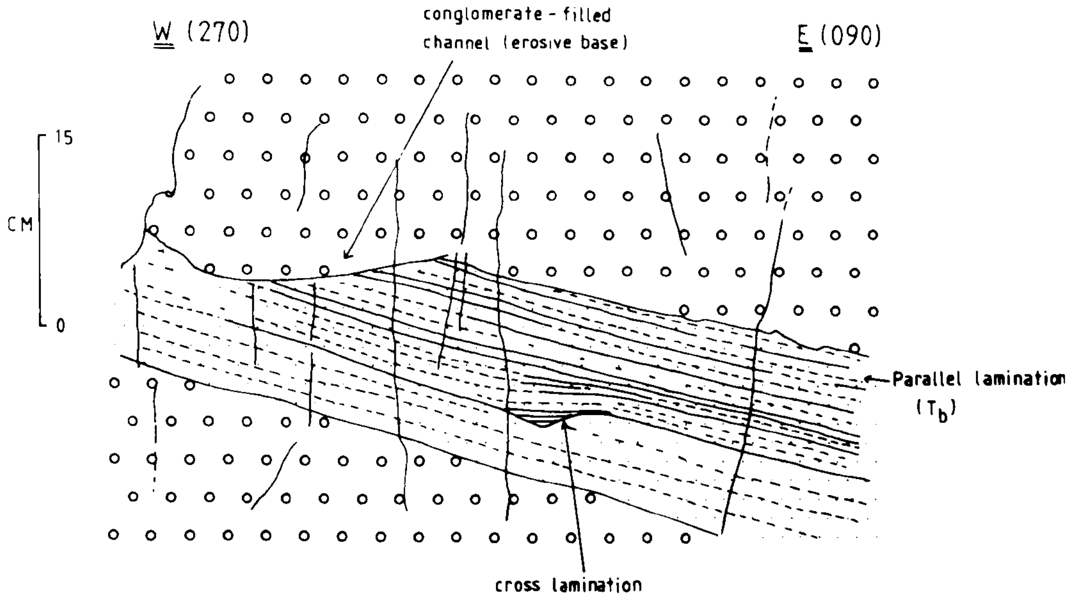


FIG 4.28 Undulatory bed boundaries, Conglomeratic Facies, Llyn y Ffran.

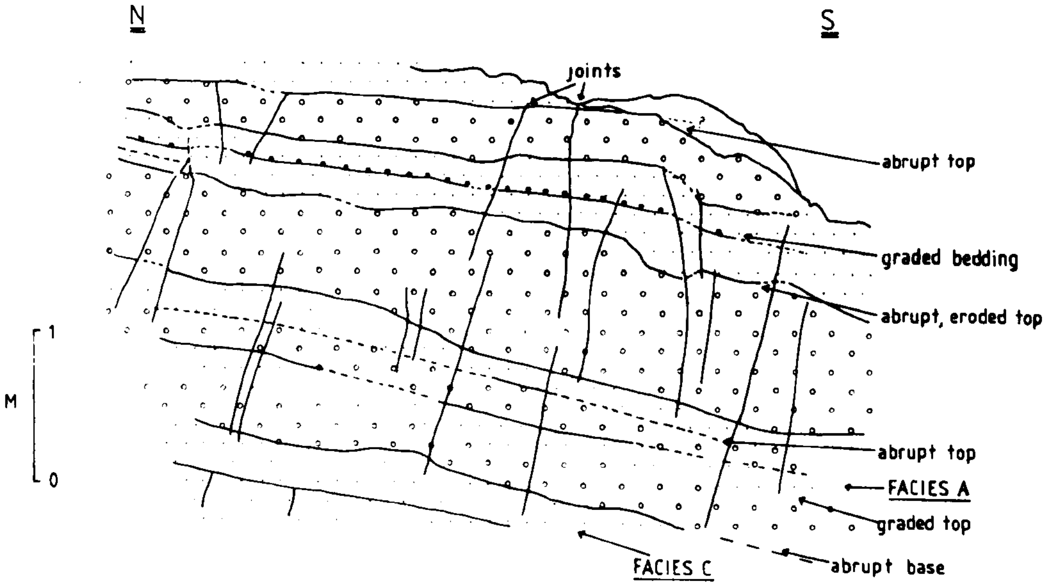
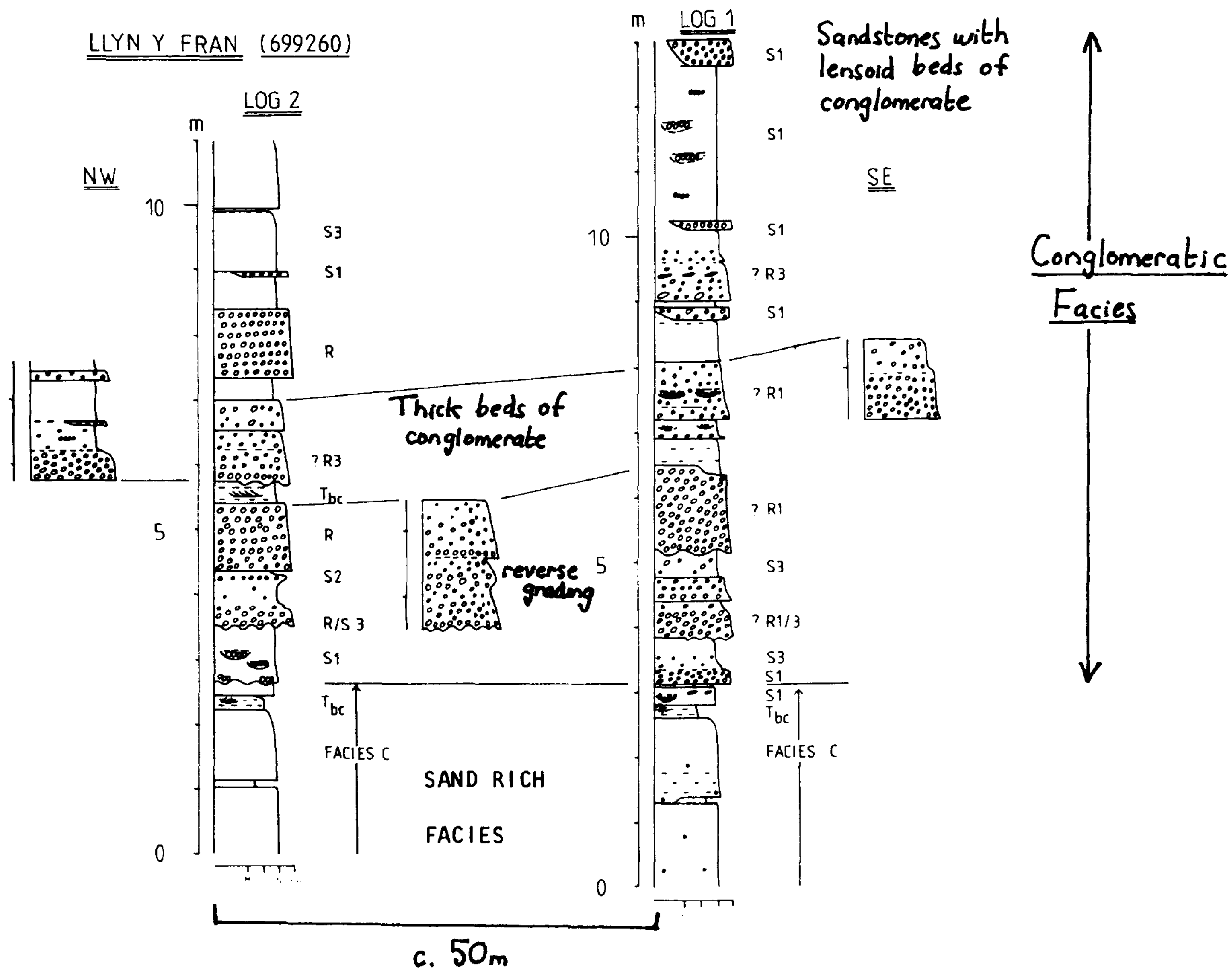


FIG 4.29 Lateral bed continuity, Conglomeratic Facies, Llyn y Ffran. using terminology of Lowe (1982)



of beds in the Conglomeratic Facies. The thicker conglomeratic beds tend to be slightly more laterally continuous than the thinner beds though the former show lateral changes in sedimentary structures e.g in Fig 4.29 at the base of the Conglomeratic Facies from S2 to S3 of Lowe (1982). The thinner conglomerate lenses usually wedge out laterally within 2m though the lenses often tend to occur preferentially at certain horizons. In general the conglomerates in the lower part of the sequence tend to be thicker bedded and more laterally continuous than the thinner lenses which dominate in the upper part of the sequence.

Vertical Sequences.

Overall the sequence at Llyn y Ffran fines and thins upwards. The lowest conglomerate has an abrupt base, resting on Sand-rich Facies. Conglomerate beds often have abrupt tops, but the sequence as a whole fines up from thick, massive conglomerate beds into sandstone beds containing thinner lenses of conglomerate (scour fills) and finally into Sand-rich Facies (conglomerate poor) containing abundant low angle cross-bedding.

Process Interpretation.

This facies resembles facies A1 of Mutti (1979) since it is dominated by conglomerate and sandstone beds which are often several metres thick and contains erosive bases, overall normal grading, occasional pebble train laminae and intraclasts. Sequences are predominantly made up of T_a beds, though the bed at 5.5m on Log 2 (Fig 4.29) can be assigned to the parallel laminated T_b division. Many of the T_a beds may be subdivided further (Lowe 1982):

i) Thick conglomerate beds with scoured bases (R1).

ii) Conglomerate filled scours within sandstone (S1).

iii) Normal graded units (R3 and S3)

iv) Rare inverse grading (S2).

Most coarse grained scour fills have abrupt tops which suggests that different grain sizes behaved differently within the flow which transported them. Lowe (1982) argues that the rudaceous grain size population behaves differently from the arenaceous component prior to deposition. The rudaceous fraction was supported by dispersive pressure and buoyant lift from upward moving fluid and was affected by basal traction. The arenaceous fraction however was dominated by coarse to medium grained sand which remained in suspension longer, due to hindered settling effects. These scour fills therefore fit Lowe's (1982) model of high density turbidites.

In other cases however coarse grained deposits pass up gradationally into finer deposits; in this case both were probably deposited relatively continuously from suspension. Thus under certain circumstances grains of different grain size populations behave in a similar way, for instance if the flow decelerates rapidly.

The Conglomeratic Facies was deposited from flows of considerable velocity and competence. The occurrence of conglomerate lenses within sandstone beds and not just at the base of sandstone beds suggests short periods of increasing flow competence produced by unsteady, surging flows. The bimodality of grain sizes within the sequence between conglomerate sized grains and medium sand possibly reflects an original supply of bimodal sediment which subsequently segregated within the flow. The poorly laminated nature of many of the beds indicates deposition too rapid to form clear lamination. The variability in thickness of the conglomerates may also reflect variation in the supply of coarse grains within the flow, though erosion may also be an important contributing factor. Sandstone beds may represent periods of lower flow velocity and locally lower rates of deposition, as indicated by the T_b beds.

Facies 3 : Amalgamated Coarse Grained Facies.

Main Characteristics.

The Amalgamated Coarse Grained Facies (AC on the logs) is characterised by thick sandstone beds up to 5m thick, most commonly between 80 and 150cm thick. This facies is sandstone dominated with sandstone-siltstone ratios usually exceeding 20. Siltstones mainly occur as thin discontinuous interbeds between the sandstones. Most thick beds are composed of coarse to medium sandstone. Clasts ranging from very coarse sand to pebbles also occur either:

a) dispersed throughout the bed (Plate 4/VI) which indicates buoyancy/grain support processes in a high density flow and/or very rapid deposition.

b) as part of the graded basal part of a bed.

c) as the infill to scours. Scours may occur either at the base of, or more rarely within, an amalgamated bed.

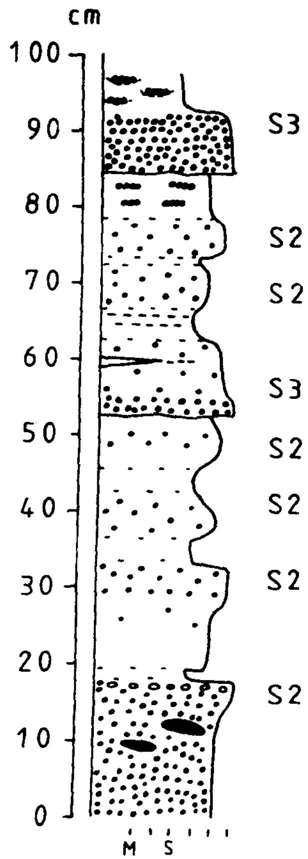
d) as diffuse patches or laminae.

Thick beds of conglomerate are relatively rare; the thickest beds are usually composed of medium sandstone.

Sedimentary Structures.

Graded bedding may be present or absent near the base of beds. Coarse tail grading is most common but distribution grading also occurs (the S3 division of Lowe 1982). Crude grading may also occur near the tops of beds and may occur rapidly or gradationally. Inverse grading may also be present (S2 division), occurring as diffuse bands (traction carpet deposits?) or more rarely as part of a general coarsening upwards of the bed (e.g F Rock, section 4, unit ii). Multiple grading is also relatively common with complex sequences of normally and inversely graded layers (Fig 4.30), which may indicate the effects of surging flow within an individual turbidity current. Lateral variations in grain size are also common, indicating substantial variability in

FIG 4.30 Log showing multiple grading, Amalgamated Coarse grained Facies in the Barmouth Formation, Bwlch y Llan.



normal and reverse graded sandstones and conglomerates.

(using terminology of Lowe 1982)

Structures in the Amalgamated Coarse grained Facies:

FIG 4.31 Rhinog Formation, Rhinog Fawr.

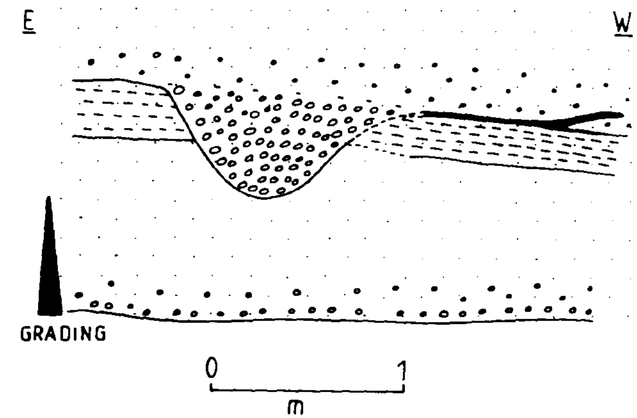


FIG 4.32 Barmouth Formation, Bwlch y Llan.

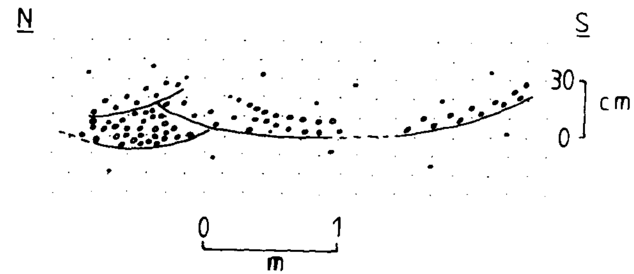


FIG 4.33 Rhinog Formation, Rhinog Fawr.

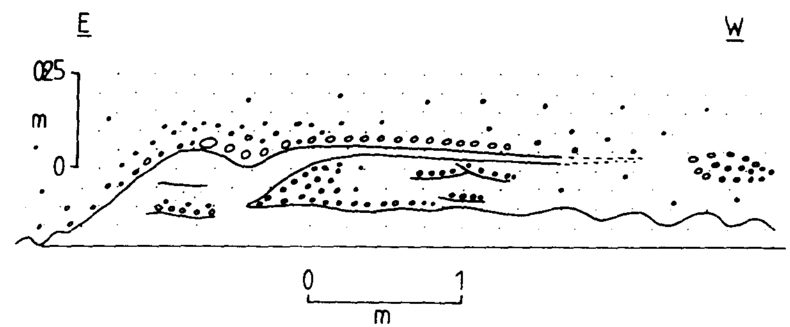


PLATE 4/VI : Dispersed grains, Amalgamated Coarse grained Facies, Rhinog Formation, above Llyn Du.

PLATE 4/VII : Multiple grading, Sand rich Facies, Barmouth Formation, Y Garn.

flow competence and/or patchy supply of coarse grained sediment.

Scour and fill structures are abundant in this facies; they may be up to 1m deep, though usually do not exceed 50cm in amplitude, with an average of 20-40cm. The scours are typically undulose and sometimes truncate bed boundaries (Fig 4.31). Scours may have steep sides though more commonly they form wide, shallow scours which in some cases truncate other scours in an irregular fashion (Fig 4.32). The scours are usually filled with coarser sediment fill than the underlying and overlying sediment. Fills of finer grained sediment than the underlying sediment were found only rarely. The fill usually contains grains ranging between coarse sand and pebble grade.

The scour fill may be of two main types:

a) graded fill, passing up gradationally into finer sediment above. Where the scour fill is only slightly coarser than the surrounding sediment the fill is often multiple graded.

b) Poorly graded to ungraded scour fill. They often occur as isolated lenses of coarser sediment.

Crimes (1970a) showed that scour and fill structures ("washouts") have variable alignment but there were frequency maxima in the W-E and NNW-SSE orientations.

Lateral Variation.

It is difficult to follow individual beds laterally because of the irregularity of the thin siltstone interbeds, though some sandstone beds appear to be lenticular at outcrop (Fig 4.33), especially the coarser beds. Sandstone packets tend to be tabular on the outcrop scale (see section 4.7).

Vertical Sequences.

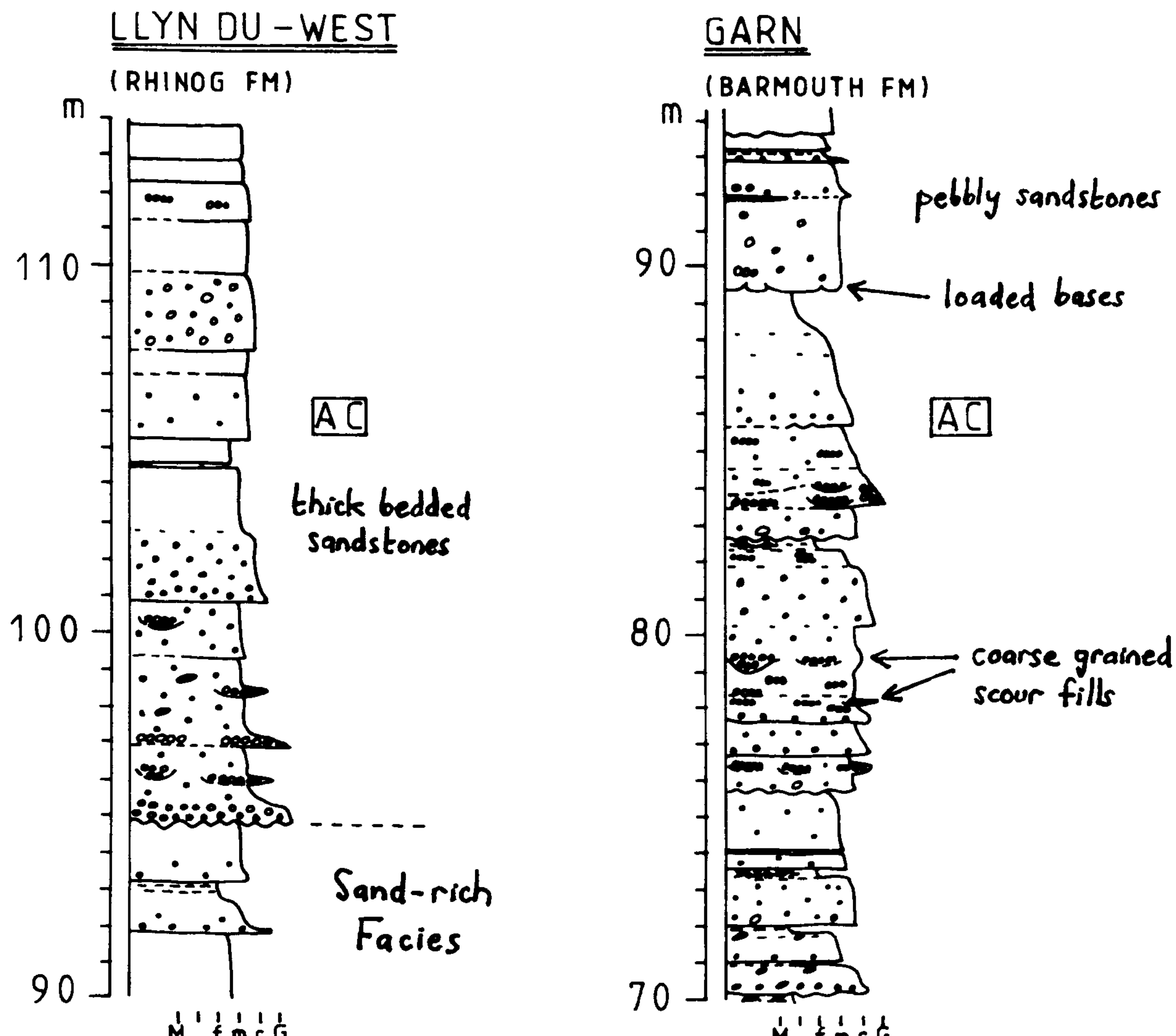
This facies often has sharp lower boundaries but transitional boundaries may occur with the Sand-rich and

Conglomeratic Facies. There is little vertical organisation within sequences. Logs typical of the facies are given in Fig 4.34.

Process Interpretation.

The Amalgamated Coarse Grained Facies often contains beds with a bimodal grain size distribution. In some beds the two grain size populations were deposited together either as a mixture of granules and sand or the granules grade up into the sand. Some coarse scour fills however have relatively abrupt tops indicating that some of the coarser grained sediment was segregated within the turbidity current at some stage prior to deposition (in a similar way to the Conglomeratic Facies). Coarser grains would tend to concentrate near the base of the flow when the flow velocity fell below the threshold velocity to keep some of the grains in suspension. The coarse grains were probably transported as traction load, but may have become segregated from the main body of the flow. Local vortices in the head of the flow would have produced scour pits which would be preferential sites of deposition (lower velocity areas) for these traction loads, so coarser grained sediment would tend to deposit in the scour pits as a lag. Sharp tops to scour fills would imply differentiation of at least two grain size populations which may have been enhanced by a lack of intervening grain sizes supplied to the flow. Where tops of scour fills are graded this suggests less complete segregation. A similar model has been proposed to account for basal lag breccias at the base of ignimbrites (Druitt & Sparks 1982). However in this model the large density contrasts between different products of pyroclastic flow deposits allow much more complete segregation within the flow. Where scour fills occur below thick massive beds this probably indicates waning flow, whereas multiple scour fills within thick beds possibly suggest surging flow or variability of supply of coarse grained detritus, spatially within the flow and/or temporally.

FIG 4.34 Example logs from the Amalgamated Coarse grained Facies.



Most beds in the Amalgamated Coarse Grained Facies belong to the Bouma T_a division and less commonly T_{ab}. Amalgamation is abundant and the thick silt interbeds (T_{de}) are uncommon. Defining the deposits of individual turbidity currents in amalgamated sequences and distinguishing them from surging flow deposits (showing evidence of repeated fluctuations in flow velocity) is difficult. Turbidites may amalgamate:

a) as a result of deposition of successive beds over a short period of time, i.e without finer grained sediment having time to deposit from suspension.

b) as a result of sediment bypassing, so fine grained sediment was deposited either distally from this site as in the case of a lobe environment, or in the case of a channel environment as overspill into the interchannel areas (in the sense of Walker 1978).

c) by erosion of the finer grained sediment by the next turbidity current, in which case one would expect to see abundant intraclasts. Intraclasts are probably not common enough to allow this to be the only cause; erosive bases are abundant however.

d) a combination of the above, which is probably the most likely.

Facies 4 : Sand-rich Facies.

Main Characteristics.

The Sand-rich Facies is quantitatively the most important facies in the Rhinog and Barmouth Formations. In the Rhinog Fawr section for example Sand-rich Facies forms about 60% of the sequence (5% Thin bedded Facies, 35% Amalgamated Coarse Grained Facies). It can be distinguished by the predominance of thick sandstone beds, which usually contain Bouma T_a units. Sandstone-siltstone ratios are usually greatly in excess of 1. The Sand-rich Facies can be distinguished from the Conglomeratic Facies by the lack of thick (metre scale) conglomerate beds. In general bedding is better developed and sequences are more organised and laterally continuous than the Amalgamated Coarse Grained Facies.

This facies includes beds ranging in grain size from pebble conglomerate to mudstone, though beds of fine to medium sandstone are the most common. The Rhinog Fawr section (from [SH 663 297] to [SH 657 292]) will be taken as an example of a sequence through the Rhinog Formation since it is the longest section that was studied in detail and is typical of the Rhinog Formation as a whole.

Beds range from 1cm to several metres thick. However beds between 20 and 90cm thick are most common and make up the largest proportion of the sequence. Beds less than 20cm thick are relatively common and similar to bed types in the Thin Bedded Facies, but they form only a small proportion of the whole sequence. Beds greater than 200cm thick are relatively rare and the maximum bed thickness is 650cm. These very thick beds are probably amalgamated and recognition of individual turbidite events is not always possible. Most graded beds (T_a) are greater than 10cm thick and most are over 30cm thick.

Since beds of approximately similar thickness often tend to occur interbedded together the Sand-rich Facies can be arbitrarily divided up into bed types on the basis of sandstone bed thickness:

i) Thin bedded 10-30cm.

ii) Medium bedded 30-70cm.

iii) Thick bedded 70-100cm.

iv) Very Thick bedded 100-250cm.

v) Very Thick Massive Beds 250-650cm.

It is also possible to classify beds on the basis of good or poor grading and on the types of Bouma sequence displayed. Subfacies can be distinguished but are rarely sharply defined since there is often considerable overlap of bed types. In general coarser grained beds or beds interbedded with coarser beds tend to be thicker bedded, whereas finer beds tend to be thinner bedded.

It is important to look at the different sedimentary structures which occur in this facies in order to determine the nature of the processes active at the time of deposition of this facies.

Graded bedding.

Many sandstone beds in the Rhinog and Barmouth Formations show graded bedding. Most of the grading types of Kuenen (1953) and Dzulynski & Walton (1965 p171) are present; discontinuous and delayed grading being the most common. In general thinner beds tend to show gradual distribution grading, whereas thicker beds are usually discontinuously graded. Thicker beds may show well defined grading near the base, be ungraded in the central part of the bed and then fine rapidly near the top. Poorly graded beds are locally common and usually only show a slight fining near the top. In terms of textural changes associated with grading, coarse tail grading is the most abundant near the base of beds, though distribution grading is also common and often predominates in the upper parts of beds (Pettijohn 1957).

Normal grading is the most common grading type, though several coarse grained beds show a basal inverse graded layer (S2 of Lowe 1982). Inverse grading tends to occur on the small scale (1-10cm scale) and is probably produced by traction carpet affects as described by Lowe (1982). Inverse grading may also occur more rarely on the larger scale (about 30cm) e.g. F Rock, section 3. Larger scale inverse grading may also be produced by traction carpets or more likely by surging flows containing irregularly distributed coarse bedload.

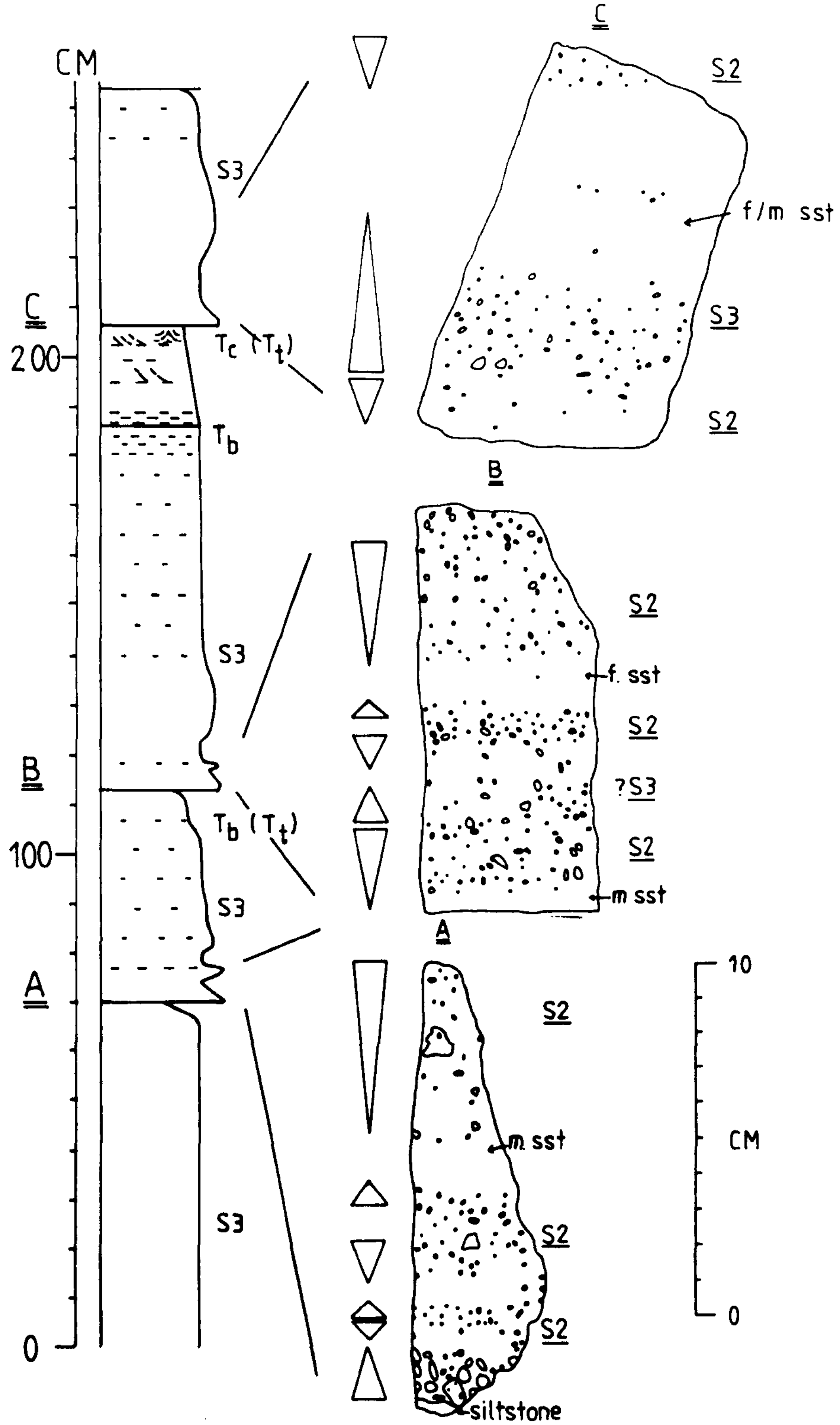
Multiple grading is also quite common and may include both normal and reverse patterns (Plate 4/VII). It is most commonly associated with coarser and thicker sandstone beds e.g Bwlch Tyddiad [SH 659 299] (Fig 4.35). They contain S2 and S3 divisions (Lowe 1982) and show lateral changes to other divisions including S1.

Thin diffuse laminae also occur within many T_a units which are often overlain by more prominent parallel laminated sandstone (T_b).

Graded sequences often show complex vertical and lateral variations in suspension dominated sedimentation (S3) and traction dominated sedimentation (S1, S2 and T_{bc}).

Grading may be gradual or relatively abrupt, associated with grain size changes either near the base or tops of beds. Some beds show subtle changes in grain size which have caused cleavage to refract. Graded beds are often very faintly laminated and some coarser beds contain alternating coarser and finer laminae which generally fine upwards, resulting from an interaction of suspension and traction deposition. Abrupt grain size changes are most common in middle absent Bouma sequences where graded T_a divisions are commonly overlain by much finer grained T_c divisions. There is often a large grain size gap between the two divisions. These characteristics indicate either distinct grain size gaps in the population of transported grains or relatively sharp segregation of grains within the flow. Segregation of grains may result in erosion or non-deposition of the intermediate grain sizes, which may be transported further and thus deposited more distally in the flow. It is also possible that rapid grading may also result

FIG 4.35 Detailed log through multiple graded Sand rich Facies, Bwlch Tyddiad (near the Roman Steps).



from differentiation of a flow into high and low density fractions.

The abundance of coarse tail grading may also indicate that there was some grain size segregation within the flow. A relatively gradual reduction in flow from a well mixed range of grain sizes (i.e. a highly turbulent flow), would produce distribution grading. However finer grained matrix is often present near the base of beds; this may have been deposited within a high density layer at the base of the turbidity current. Grading restricted to the coarse grained fraction of grains implies rapid sedimentation since only the coarsest grains have time to fall through the flow. Other grains are supported by dispersion and other buoyancy affects and thus remain in suspension. Coarse tail grading also indicates either incomplete turbulent mixing of the sediment or grain size gaps in the sediment being supplied.

Sole structures.

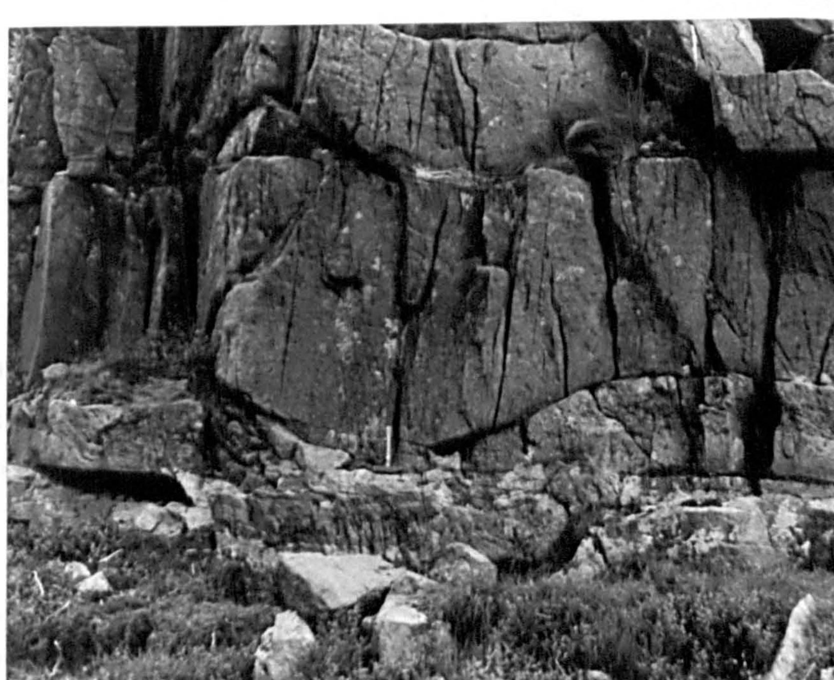
Load casts are particularly common in the Rhinog and Barmouth Formations. They are often associated with sand or mud injection from the underlying bed. The loads have a variable wavelength and amplitude and are highly irregular in shape, from bed to bed or even laterally at the same horizon. Some however are aligned parallel to flute/groove casts and may result from loading of a pre-existing flow-produced sole structure. In some places there is a consistent direction of flame injection which probably also suggests an original flow origin.

There may be variable amounts of down-loading, from undulose loaded bases to isolated "pods", usually of coarse sandstone (originally small scour lags) which were down-loaded into the bed below. They are often underlain by siltstones, which have compacted around these lenses. Transitional between the two end-members there are scour fills which have down-loaded and may thicken and thin laterally.

Scours are the main type of flow-produced sole structure (e.g. Plate 4/VIII); these range from a few

PLATE 4/VIII : Conglomerate filled scour, Sand rich Facies, Rhinog Formation, Rhinog Fawr.

PLATE 4/IX : Large scour, Sand rich Facies, Rhinog Formation, west of F Rock.



centimetres to about 100cm deep and may be tens of centimetres to several metres wide. They may have a coarse fill (up to pebble conglomerate). Scours may occur at the base of beds (e.g Plate 4/IX) or within beds, though they are often concentrated at certain horizons, which are possibly amalgamation surfaces. Scours show variable alignments, but two maxima occur oriented approximately N-S and E-W (Crimes 1970a and present observations).

Flutes are rare, though they have been recorded previously (Kopstein 1954, Crimes 1970a) and indicate flow from the north in the Rhinog Formation and from the south in the Barmouth Formation. Groove casts are more common and indicate flow in the N-S orientation. These sole structures are more common in the thinner bedded parts of the sequence and rare in the thicker bedded, coarser grained parts of the sequence where scours are more common.

Internal Structures.

Parallel lamination commonly overlies the more massive sandstones (some of which show subtle, faint lamination) as part of T_{ab} sequences. Cross lamination (T_c) is also quite common and is similar to that described in the Thin Bedded Facies, though ripple drift is more common e.g F Rock, section 1. Cross-bedding is also common in parts of the Rhinog Formation and is discussed in section 4.6.

Convolute lamination is common in the Sand-rich Facies, usually occurring near the top of sandstone packets. The convolutions typically have an amplitude of 5-15cm and a wavelength of 10-30cm. They have sharp ridges and broader intervening troughs which are aligned approximately N-S and are similar to the convolute lamination in the Thin Bedded Facies. However in the Sand-rich Facies convolute lamination generally occurs on a larger scale. Their N-S alignment suggests either W-E compression during later tectonic deformation or more likely east to west flowing currents which may have been of a similar type to those which produced the cross lamination in the Thin Bedded Facies or the cross-bedding in the Sand-rich Facies. These convolutions probably formed as a result of dewatering though there is usually complete preservation of laminae. However in some cases (e.g at Garn), lamination may be locally obliterated in the crest region of the convolution to form pipes.

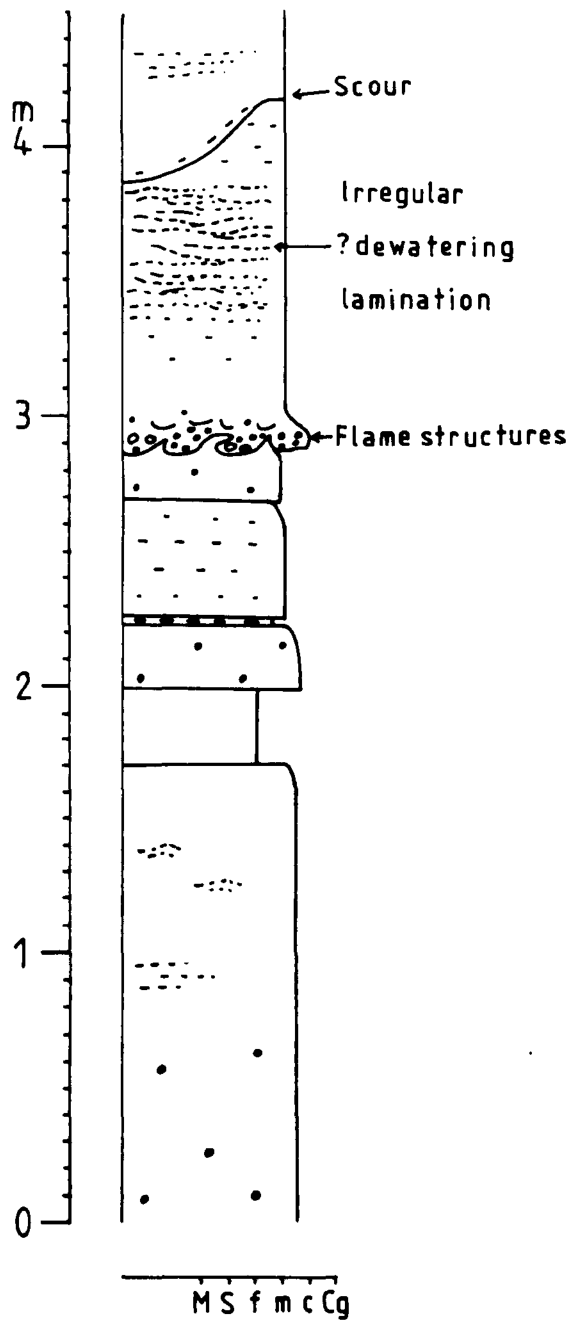
Other dewatering structures are also found. Some beds e.g near locality TF 1, north of the Roman Steps [SH 663 305] show chaotic bedding, the result of partial fluidisation of convolute lamination during dewatering, producing a complex mix of convolute laminated beds with synsedimentary sandstone intrusions. At F Rock (Section 3) a granule filled clastic dyke was produced either by downward infilling or more likely by intrusion up from a scour and fill structure into the thick cross-bedded horizon above. At locality TF 2, north of the Roman Steps [SH 665 310] a pipe structure 3cm wide and 10cm deep of unlaminated sediment occurs between upturned parallel laminae. The pipe widens near the base and was probably a region of preferential water escape.

At the Roman Steps (Fig 4.36) [SH 653 304] faint but relatively continuous (on the 10cm scale) "laminae" occur. These structures were produced relatively early since they are affected by loading. They also occur preferentially in a particular band above a flamed base and thus may be a form of lamination produced by dewatering. However they do not have the typical concave-up geometry associated with dish structures and other dewatering structures (Lowe & LoPiccolo 1974). Their association with more massive beds nearby may indicate they were both produced from rapidly deposited high density turbidites.

Intraclasts.

Intraclasts are present in many sandstone beds and vary in size from 0.5cm to 100-200cm long. They range from angular to rounded and many are approximately elliptical in shape. Intraclasts may occur at different levels within a bed though usually they are most common near the top of beds due to their greater buoyancy within the flow. The presence

FIG 4.36 Roman Steps log, Rhinog Formation.



of intraclasts indicates penecontemporaneous erosion, usually of Thin Bedded Facies type sediment, suggesting localised scour (see also F Rock, section 1).

Intraclasts occasionally form an "intraclast breccia" e.g. on the lower slopes of Rhinog Fawr [SH 6618 2970]. At this locality "intraclast breccia" directly overlies a thick sequence of Thin Bedded Facies. The lower surface of the breccia bed is erosive, though it is commonly planar, cutting down through the stratigraphy in a step-like fashion. The intraclasts are blocky and tabular and occur within a matrix of muddy sandstone. Some of the intraclasts are very large at this locality- one for instance (Fig 4.37) is 200cm long and about 15cm thick and is convolute laminated. The laminae are truncated on the lower surface of the intraclast indicating that the intraclast was probably inverted since the formation of the convolute lamination. Laterally (Fig 4.38) the concentration of intraclasts varies. The angularity of the clasts suggests that they have not been transported far and it is interesting that at this locality there is a scour on the base of the breccia bed which is of a similar amplitude to the thickness of the intraclasts. Therefore the intraclasts were locally derived, probably by erosion of Thin Bedded Facies preferentially along the bedding planes. The intraclasts show a variety of orientations and some are imbricated indicating orientation by flow. A low density, highly turbulent flow would have tended to break up the intraclasts into much smaller blocks, whereas the presence of large, angular blocks, some midway in the bed suggests that the flow which deposited them derived them locally and had a sufficiently high yield strength to support these clasts within the flow. Displacive pore pressure and grain interaction may have allowed large clasts to behave as "rafts" floating in a turbid suspension. The incorporation of large intraclasts within the flow may have resulted in a substantial reduction in turbulence and caused deposition and/or transformation to a higher density flow. The characteristics described suggest deposition from a high density turbidity current, possibly transitional with

FIG 4.37 "Intraclast breccia", Sand-rich Facies, north-east slopes of Rhinog Fawr.

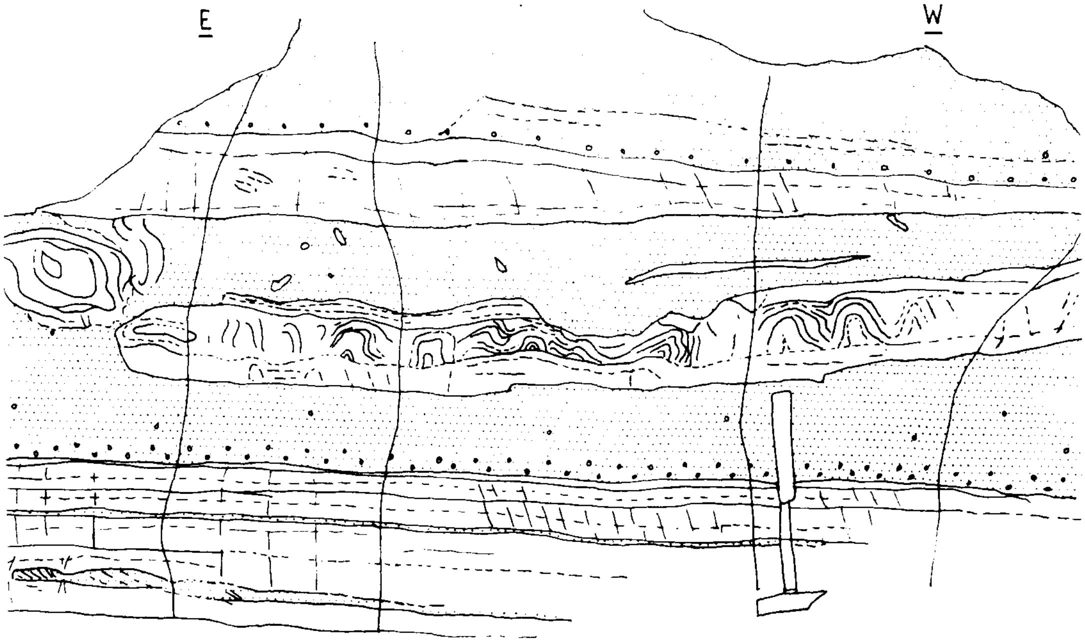
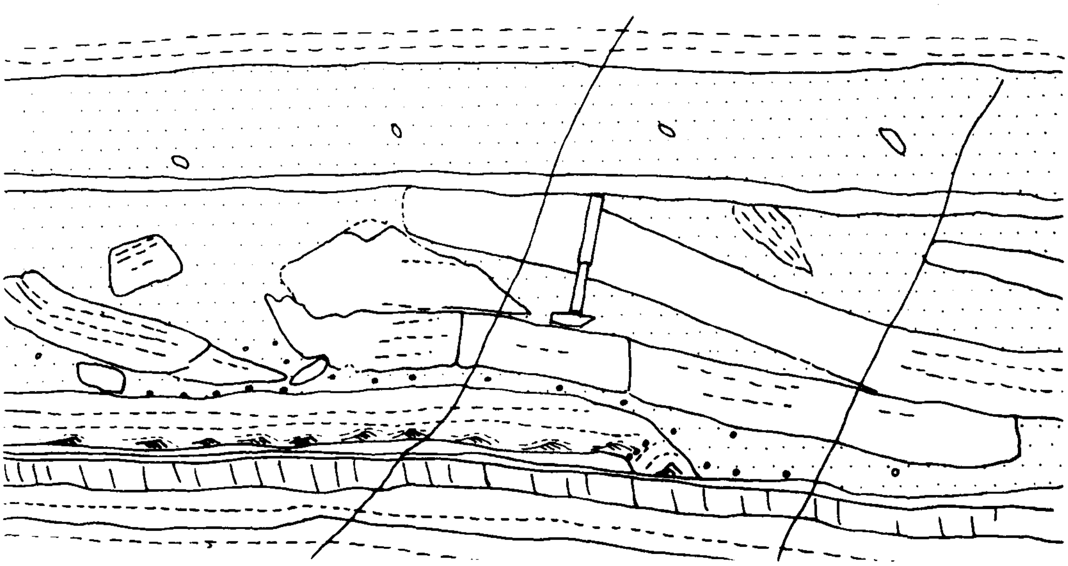


FIG 4.38 Imbricated clasts, "intraclast breccia", Sand-rich Facies, north-east slopes of Rhinog Fawr.



a debris flow. The occurrence of "intraclast breccias" is strongly indicative of deep erosion and a probable channelised origin for these beds. However "intraclasts breccias" are generally rare within the Rhinog Formation.

Vertical Sequences.

The dominant Bouma divisions in the Sand-rich Facies are T_a and T_b with minor amounts of T_c and T_d. Most beds are graded, faintly laminated or massive in their lower parts. T_b often only forms the top few centimetres of sandstone beds and generally only makes up a small proportion of Bouma sequences, especially when beds are thick and amalgamated. Thus top absent (frequently amalgamated) and middle absent Bouma sequences are the most common. Complete Bouma sequences e.g Plate 4/X are relatively rare.

As examples of the organisation and variability of the Sand-rich Facies four example logs are given from the Rhinog Fawr section (Fig 4.39):

Log 1. This log shows medium to thick bedded subfacies (50-55m), typically containing T_a sequences, with occasional T_b and an isolated occurrence of cross-bedding. This subfacies is overlain by thin to medium subfacies (55-65m) which is predominantly T_{ab} with erosive bases and a slightly coarser grain size than below. There appear to be alternations of amalgamated beds (4-5 beds in a packet) interbedded with unamalgamated ones. The thin to medium subfacies is abruptly overlain by the Amalgamated Coarse Grained Facies.

Log 2. Thin to medium subfacies occurs between 140 and 158m and includes planar based T_a units, abundant T_b, T_{bc} with convolute lamination as well as isolated scours and cross-bedding. This log contains beds of similar bed thickness to log 1, but with contrasting sedimentary structures. The thin to medium bedded subfacies is sharply overlain by Amalgamated Coarse Grained Facies.

Log 3. Overlying the Thin bedded Facies are thin

PLATE 4/X : A T_{abcde} Bouma sequence, Rhinog Formation, Roman Steps.

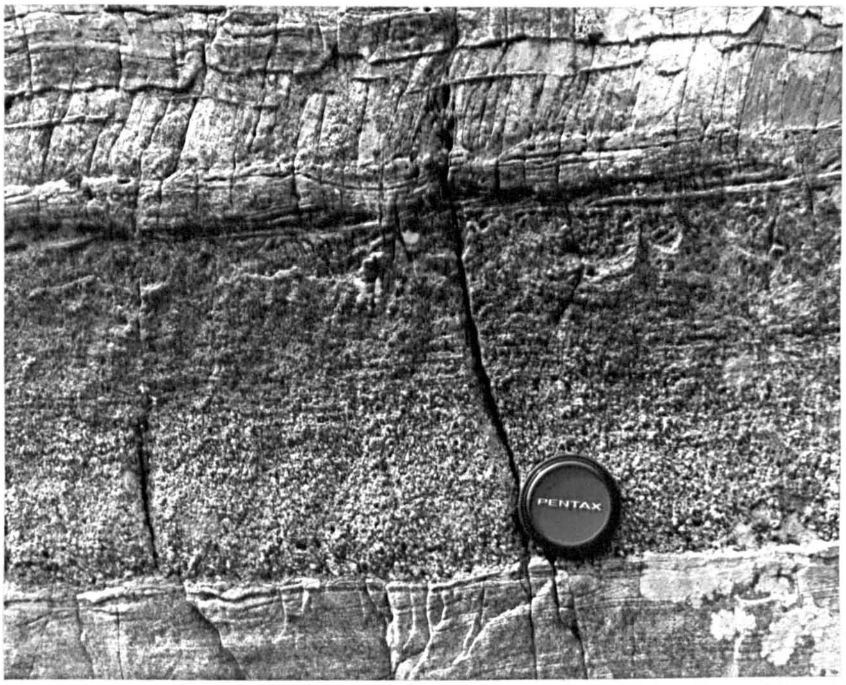
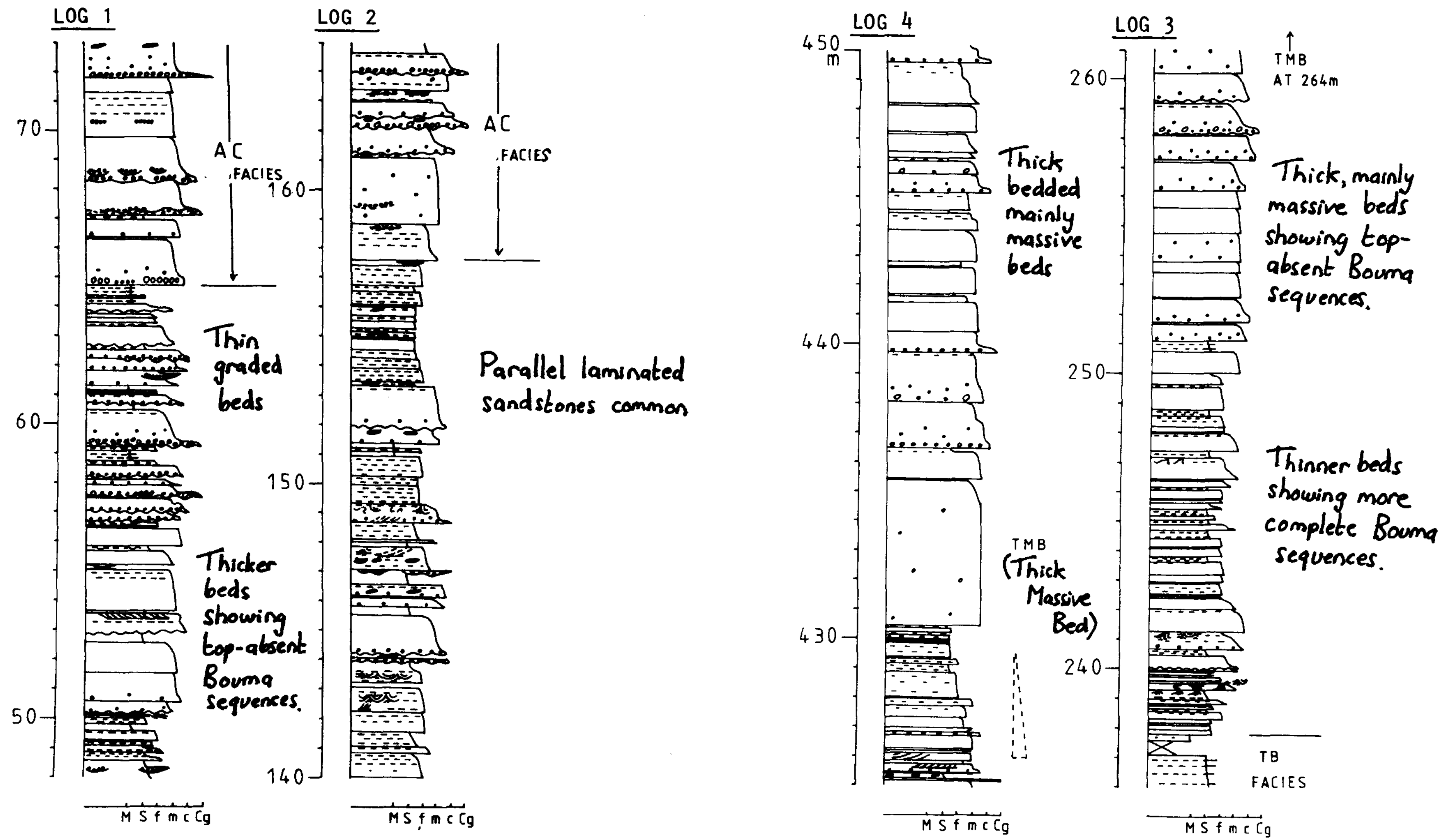


FIG 4.39 Example logs from the Sand rich Facies, Rhinog Fawr log.



beds of the Sand-rich Facies (238-246m) containing T_{ab}, T_{bc} and T_c; convolute lamination is also common. The top is transitional (246-251m) containing medium bedded, unamalgamated T_{abc} beds. Above 251m the beds are medium to thick bedded T_a and T_{ab} units with planar bases. A Thick Massive Bed (TMB) at 264m thus forms the upper part of a 25m scale thickening upward sequence. However the Thick Massive Bed is overlain by beds of similar type to those that underlie it and thus may have resulted from an isolated large-scale event. Whether there is a large scale symmetrical cycle is unclear.

Log 4. This log shows thinly bedded graded subfacies which fines upwards (425-430m) and is abruptly overlain by a Thick Massive Bed (430-435m). Above 435m coarse grained, thick bedded subfacies occurs which is transitional in type with the Amalgamated Coarse Grained Facies.

These logs emphasise the heterogeneity of this facies in which consistent organisation is lacking.

Process Interpretation.

The Sand-rich Facies contains predominantly top absent, some middle absent and only rare base absent Bouma sequences, typical of "proximal" turbidites (*sensu* Walker 1967). This facies contains beds which show abundant evidence for deposition from high density turbidity currents (S1, S2 and S3). The presence of coarse scour fills with abrupt tops as well as coarse tail grading probably indicates grain size segregation within the flow. Cross-bedding however indicates the presence of much more dilute tractional currents which may or may not be turbidity current related (see section 4.6). The abundance of dewatering structures and load structures and the general lack of mud within sequences suggests that these beds were rapidly deposited both as individual events and as groups of events.

4.6 : Cross-Bedding.

Cross-bedding is locally quite common within the Rhinog Formation, and also occurs to a lesser extent in the Barmouth Formation. Set heights are typically of the order of 10 to 30cm with a maximum of 60cm. Sets usually occur singly but more rarely as cosets incorporating up to 3 sets. Cross-bedding most frequently occurs directly above thick (c.1m) beds, though more rarely they may be interbedded with siltstones. The underlying beds are often poorly graded, occasionally with parallel lamination (T_{ab}) and there is also a common association between slightly coarser beds (e.g pebbly sandstones) and cross-bedding. The cross-bedded units are composed of grain sizes ranging from fine sandstone to granule conglomerate, most commonly occurring in medium to coarse sandstone. These deposits are usually better sorted than the underlying beds and may be coarser grained. The cross-bedded sets are commonly sharp-based and may infill scours which are erosive into the sandstone bed below. The sets also have sharp tops and are frequently overlain by thin beds of cleaved siltstone or mudstone.

Tabular and trough cross-bedding are present. They may occur as part of scour fills, though elsewhere the presence of positive topographic relief on set boundaries indicates the presence of megaripple bedforms. Palaeocurrents directions derived from the cross-bedding are variable, though they predominantly indicate flow towards the west (including flow towards the northwest and southwest).

Types of cross-bedded unit.

1) Large amplitude megaripples- for instance at F Rock [SH 6652 3107] (Fig 4.40, Plates 4/XI and XII). The cross-bedded unit exposed here has a maximum amplitude of 60cm at locality B and thins to the east (within 10m) to 30cm and to the west, lensing out completely by locality D (see section 4.7). Therefore the bedform has a minimum

FIG 4.40 Map of the F Rock area.

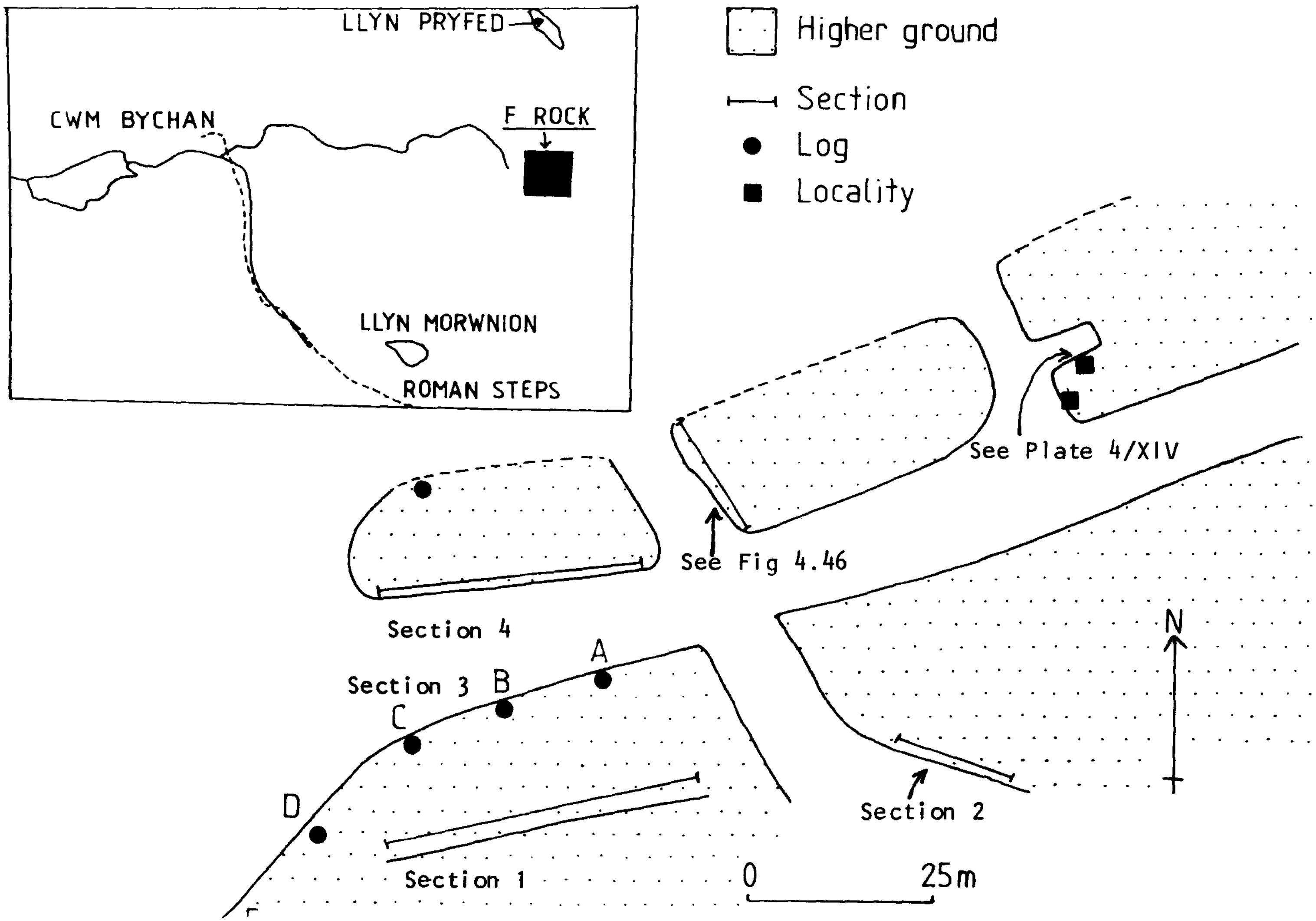


PLATE 4/XI : Large amplitude cross-bedding (type 1), Rhinog Formation, F Rock.

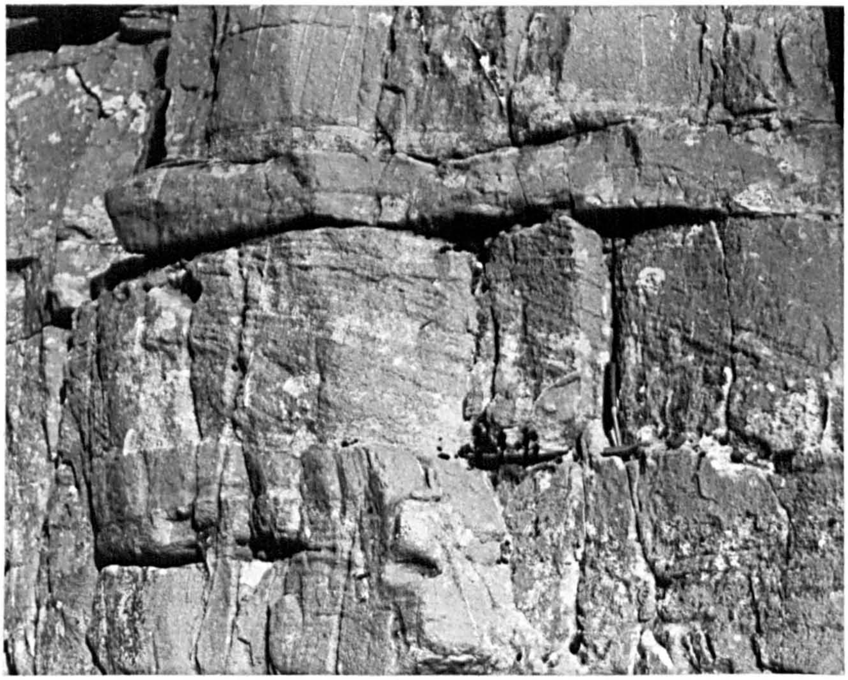
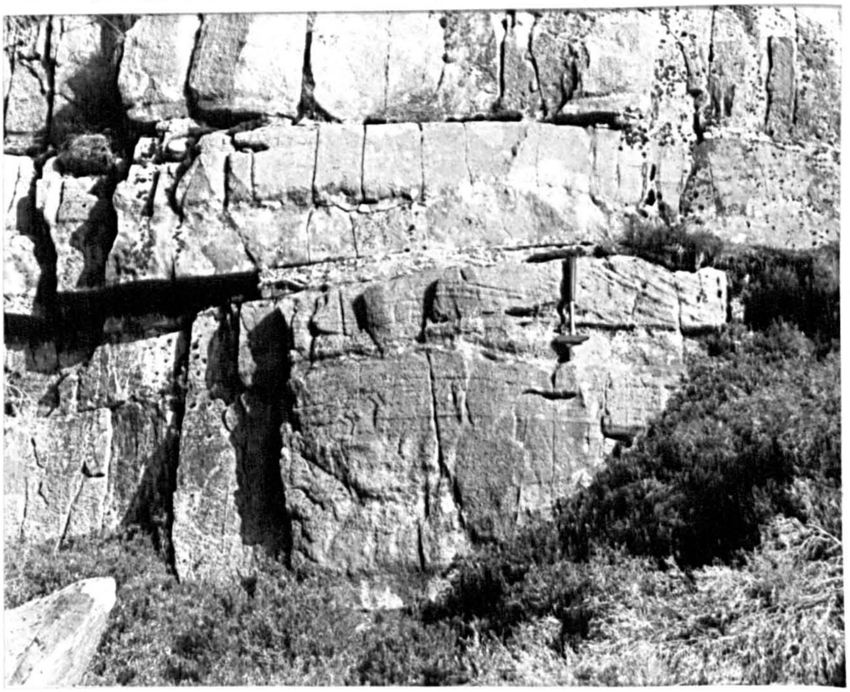


PLATE 4/XII : Cross-bedded set 10m east of Plate 4/XI.



wavelength of 30m.

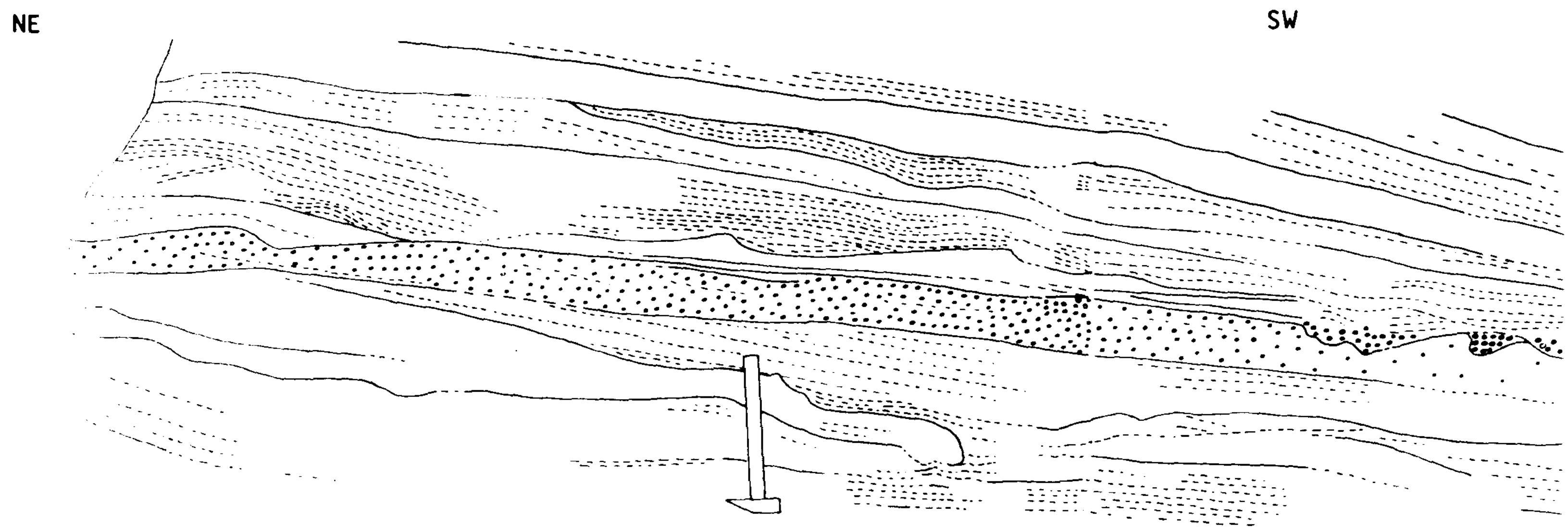
The cross-bedded unit is composed of relatively well sorted granule conglomerate to very coarse sandstone, more coarser grained than any of the grains in the underlying bed. It has a sharp base and top, the base is locally erosive, but the cross-bedding does not simply infill an erosive hollow, since there is some positive relief on the upper bedding surface, indicative of a megaripple bedform. The cross-bedded unit is made up of two sets, a thicker lower set with steeply dipping (c.20°) tabular foresets indicating flow towards the west-south-west. There are relatively angular basal contacts to foresets and tangential upper contacts with topsets preserved. The upper set is thinner (10-20cm thick), lower angle, more asymptotic (trough) and indicates flow oblique to the underlying set.

This exposure also shows interesting lateral changes in the characteristics of the bedform, particularly as one follows the lower set from east to west from locality A to B (i.e downcurrent). There is a transition from low-angle (5 to 10 degrees) dipping laminae to steeper (c.20 degree) convex-upward foresets, steepening progressively towards the west as the megaripple gradually developed.

In general large-scale cross-bedded bedforms are rare in the Rhinog Formation.

2) <u>Small amplitude megaripples</u>- for instance below Llyn Du [SH 6552 2961] (Figs 4.41 and 4.42). Here both small ripples (amplitude 5cm, wavelength c.100cm) and larger ripples (amplitude 15cm, wavelength c.75cm) are present as single sets and often overlie thick sandstone beds (Plate 4/XIII). Many show preservation of topsets and most indicate flow towards the west. Tops of bedforms often have substantial topographic relief. This type of cross-bedding is often also associated with other types of cross-bedding, in particular type 6. There is also evidence for similar current activity sculpting/scouring coarser grained beds.

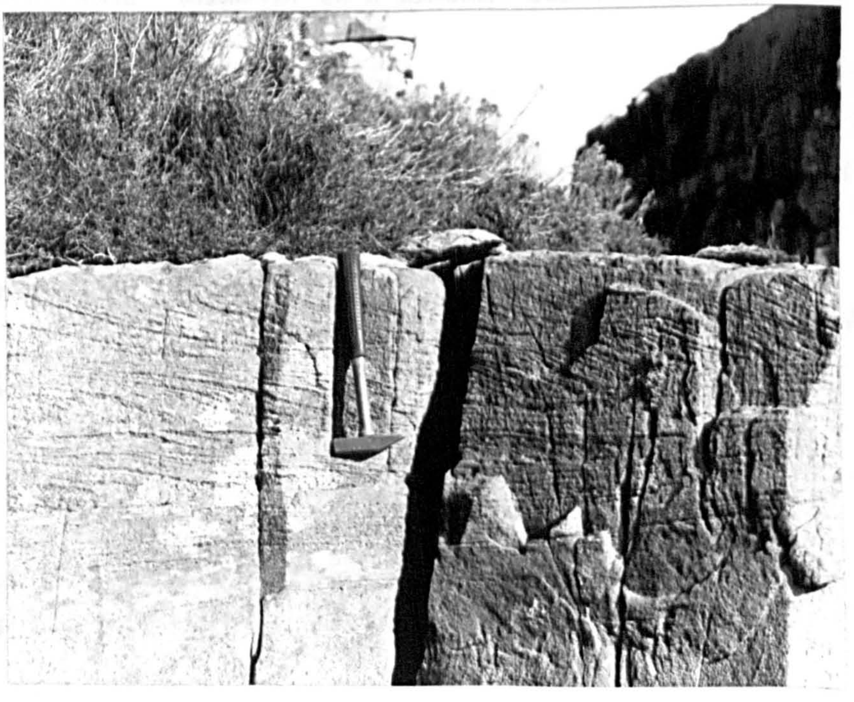
3) <u>Small scour fills</u>- for instance Y Garn [SH



FIGS 4.41 and 4.42 Small amplitude cross-bedding, Rhinog Formation, below Llyn Du.

PLATE 4/XIII : Small amplitude cross-bedding (type 2), Rhinog Formation, north-east slopes of Rhinog Fawr.

PLATE 4/XIV : Multiple set cross-bedding (type 5), Rhinog Formation, F Rock.



703 230] in the Barmouth Formation (Fig 4.43a and b). At Y Garn there is a NW-SE aligned scour, with a maximum depth of 15cm, which erodes down into the parallel laminated top of the sandstone bed below. The scour was then infilled with slightly better sorted sandstone (which weathers white), showing trough cross-bedding as a result of flow towards the southwest. As the scour filled, the foresets became progressively lower angled. Whether the scour and its fill are genetically related or whether the current which deposited the sand simply filled a pre-existing scour is unclear. However the cross-bedded unit is both underlain by a turbidite (T_{ab}) sandstone and overlain by a turbidite (T_a) showing repeated S2 and S3 divisions indicating the density current- dominated context of the cross-bedding. Similar examples are common in the Barmouth Formation e.g Garn above Barmouth [SH 618 165].

4) Large scour fills- e.g Rhinog Fawr log [SH 659 294]. The most common type of cross-bedding in the Rhinog and Barmouth Formations is the 20-30cm thick single set type with planar upper bedding surfaces. The foresets may have a tabular and/or trough geometry and may possibly represent the infills of broad shallow scours, though in only a few cases is exposure good enough to see direct evidence for this. However at a locality below Llyn Du [SH 6552 2961] (Fig 4.44) cross-bedding of 25cm amplitude occupies a wide scour in the underlying bed and so the cross-bedded set has a lensoid geometry. The cross-bedding indicates flow towards the west and when traced down-current shows syn-sedimentary folding of foresets, probably due to current-shear.

The cross-bedded units are commonly composed of slightly better sorted sediment than the bed below and may have been produced by winnowing of sediment.

5) Multiple set cross-bedding- for instance at the Roman Steps [SH 653 304] (Fig 4.45). This locality shows cosets with sets of considerably different thickness. A bed

FIG 4.43 Small scour fills, Barmouth Formation Y Garn.

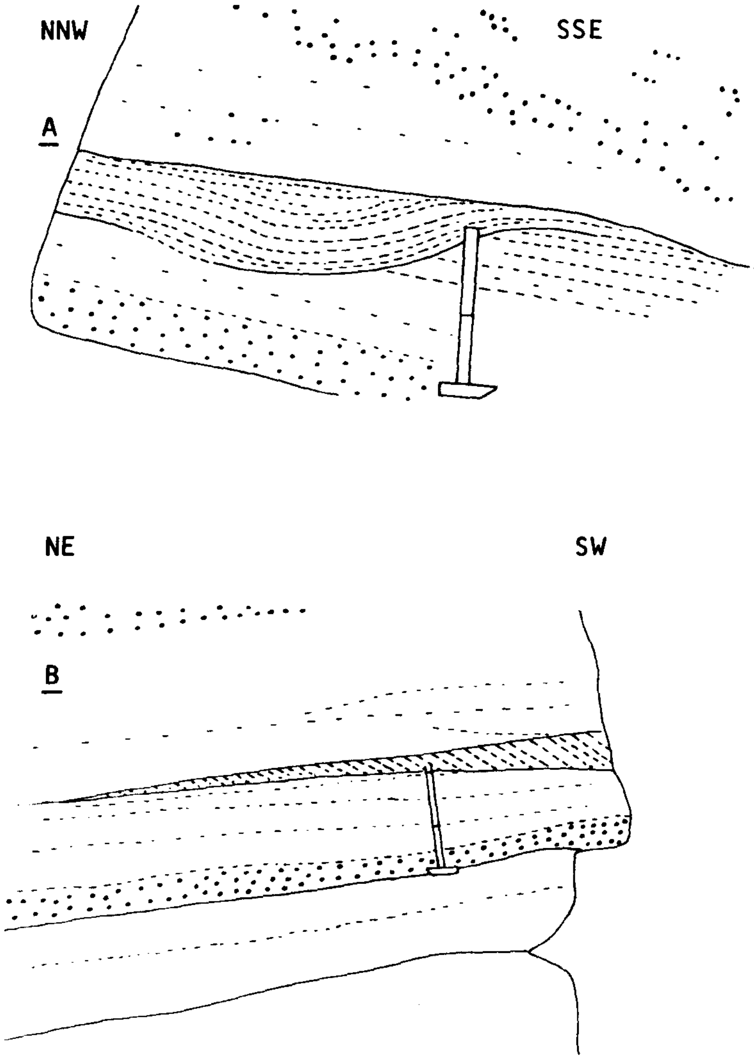


FIG 4.44 Large scour fill, Rhinog Formation, below Llyn Du.

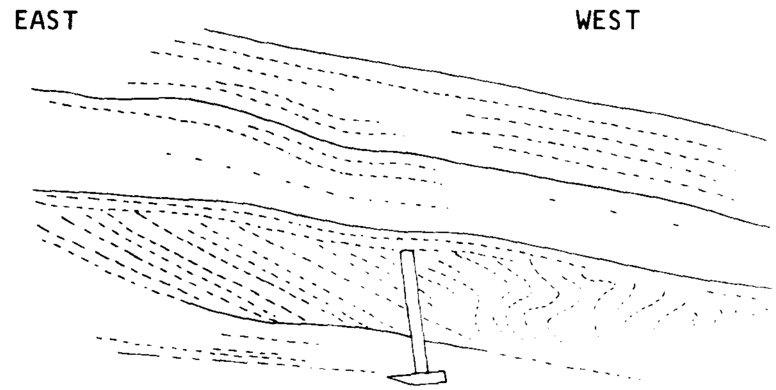
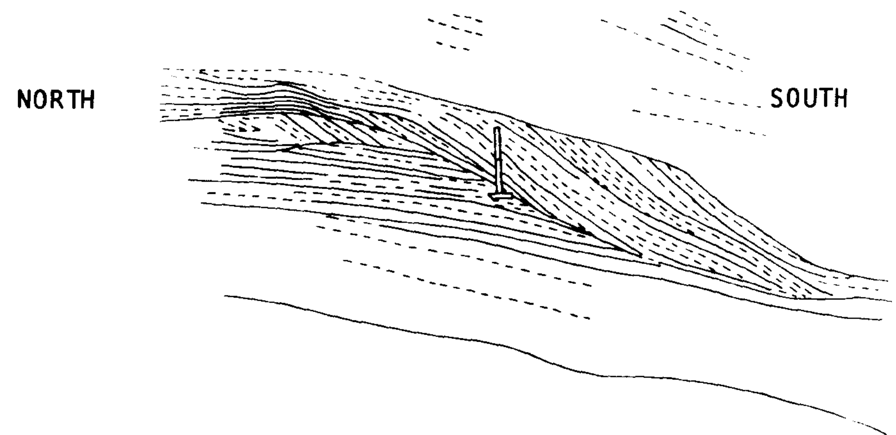


FIG 4.45 Multiple set cross-bedding, Rhinog Formation, Roman Steps.



of parallel laminated sandstone is overlain by a cross-bedded set 10cm thick, the latter eroding slightly into the former. These two beds are then overlain and eroded deeply before the deposition of a set of larger scale cross-bedding. As one traces this set from north to south the set thins slightly over the crest of the underlying bedform and then the foresets steepen (dipping towards the south-south-east) and migrate down the margin of the scour so that the set thickens to about 40cm. The cross laminae have relatively tangential bases and sharp, angular tops due to erosion of the tops of the foresets prior to deposition of the overlying thick-bedded, faintly laminated sandstone. This contact is erosive and includes a scour which is approximately parallel to the foreset dip, though some cross laminae are truncated. The scour surface may represent the approximate down-current limit of the bedform or may represent a cross lamina that was preferentially exploited by the eroding current. The similarity in grain size and the amalgamated nature of these deposits suggests that they were deposited by flow events (probably separate events) which were closely connected by similar processes over a relatively short period of time to produce composite bedforms.

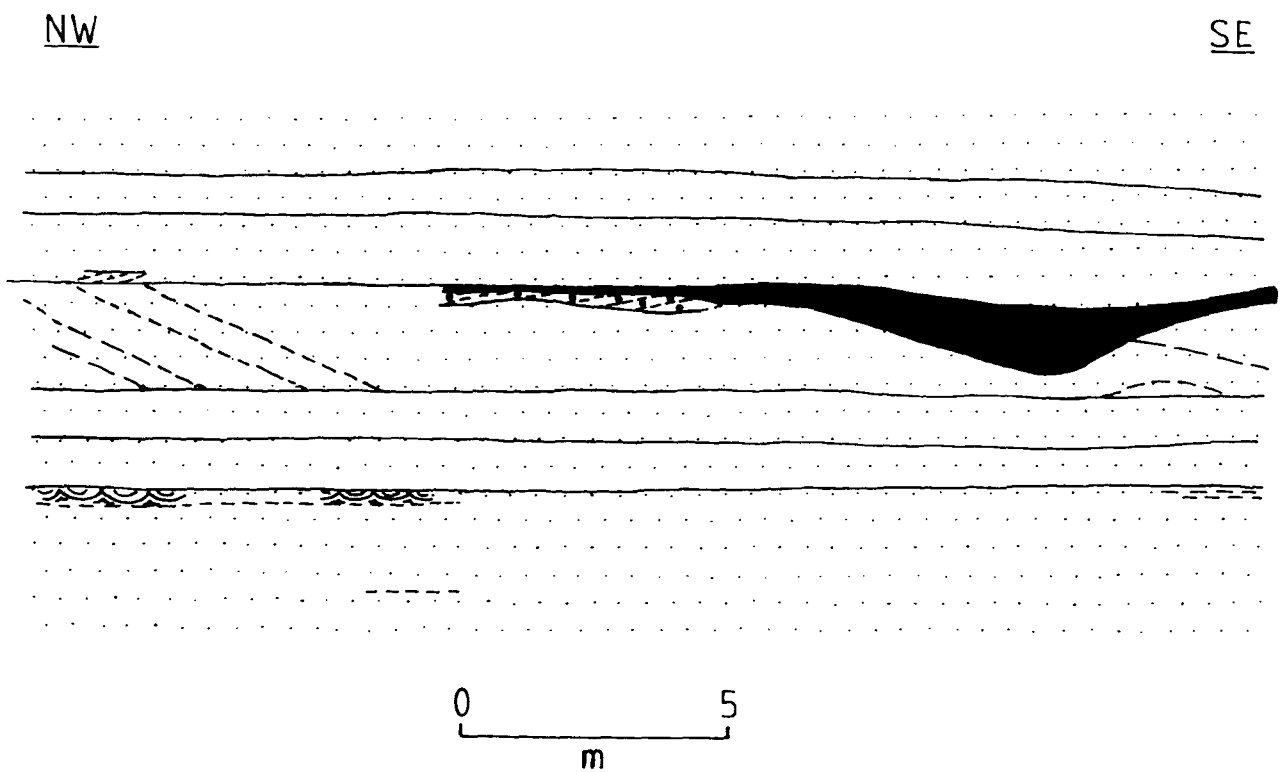
At F Rock, locality 1, the sequence overall coarsens upwards over a few tens of centimetres (Plate 4/XIV), which is unusual within turbidite sequences. Faintly laminated medium sandstone is overlain by parallel laminated coarse sandstone. This is in turn overlain by a set of tabular cross-bedding which has a maximum thickness of about 10cm. The cross laminae dip consistently towards the west and the set also thins in this direction. The cross laminae are gently convex upwards and cannot be sub-divided into bottomsets, foresets and topsets but show angular tops due to erosion. The set above begins (in a downcurrent direction, i.e. east to west) with low angle cross stratification and cross-beds similar to the set below. The cross laminae are unusual in that they have angular bases and tangential tops, are gently convex upwards and generally thicken down-dip. The lamination steepens downcurrent and there is a rapid transition from the type of bedform described above to one where clear foresets and topsets can be distinguished. The new bedform also climbs at about 20 degrees towards the west and indicates flow towards the west-north-west. These features indicate progressive bedform aggradation in response to an increase in the amount of sediment coming out of suspension. The sediment fill occurs within a large trough as seen in an exposure at right angles to the current. Locally the upper part of the fill shows oversteepened foresets and synsedimentary folding produced by current shear, probably towards the north-west.

6) Low angle cross-bedding- for instance F Rock. This type of cross-bedding is very variable, usually with a dip of less than 10°, though it may form consistently dipping foresets or isolated draping of scour surfaces that are part of complex amalgamated scour and fill events. They do not usually constitute discrete bedforms, though the association of low angle cross-bedding and apparently parallel lamination may possibly suggest the presence of low amplitude sand waves similar to those described by Smith (1971) in Recent fluvial deposits in the River Platte, U.S.A.

7) Large scale diffuse cross-bedding- for instance at the Roman Steps [SH 653 304]. Rarely one may observe metre-scale, diffuse dipping surfaces in otherwise thick bedded, superficially massive beds. These surfaces often have a variable direction of dip and few consistent laminae dipping parallel to it. Therefore they do not seem to form true bedforms.

8) Large scale cross-bedding with graded cross beds- for instance F Rock, locality 2 (Fig 4.46). This locality shows cross-beds with a set height of 167cm, dipping towards the southeast with individual graded, tabular cross-beds between 48 and 90cm thick, very similar

FIG 4.46 Large scale cross-bedding, Rhinog Formation, F Rock.



to many of the underlying horizontally bedded turbidite sandstones. This set has abrupt upper and lower contacts and passes laterally to the northwest and southeast into massive sandstone, though further to the southeast cross-bedding appears, also indicating flow towards the southeast. This very large bedform(?) is different from the cross-bedded units described above. The thickness of individual massive cross-beds indicates very large scale migration events, successively dumping turbidite sand on the lee-side of a large scour, without significant amounts of avalanching. This bedform is however problematical in that it requires the rapid dumping of turbidite sand at a depositional angle of 20°; one might expect instead horizontal infill of so large a scour, especially in the absence of other smaller scale tractional features.

This bedform or scour fill was subsequently reworked and eroded, producing more typical cross-bedding (type 2) with a set height of up to 34cm, but indicating flow towards the northwest (i.e. opposed to the larger scale cross bedding beneath). There was then a period of erosion, producing a scour (N-S aligned), which was then mud-filled. Thus this sequence represents the complex interplay of different tractional and non-tractional processes.

Process Interpretation.

Cross-bedding in the Rhinog and Barmouth Formations indicates deposition by westerly-flowing, traction-dominated currents. These currents either infilled scours (types 3 and 4) or produced megaripple bedforms (types 1 and 2). The common preservation of topsets suggest rapid deposition without large amounts of erosion, often associated with faint cross laminae. Other sets however show relatively well sorted, alternating coarse and fine laminae, with angular, truncated tops to foresets indicative of slightly slower, more intermittent deposition. The lack of climbing bedforms and the angular lower contacts of many of the cross laminae suggest comparatively little sediment fallout from suspension, indicating deposition from the bedload of a relatively dilute flow. The presence of low angle cross-bedding possibly indicates periods of high energy flow, while the presence of large bedforms and cosets indicate sustained periods of flow, with flow velocity values in the megaripple field of the bedform stability diagram (e.g. Allen 1970, 1982).

If we take the large bedform at F Rock (see type 1) as an example, we may use its characteristics to help confine the flow characteristics of the current that produced it. The bedform is composed of granule conglomerate (average grain size 2mm) and megaripples of this grain size form at flow velocities of between 60 and 150cm/sec (Middleton & Southard 1977). Given the fact that the maximum waveheight of the bedform is 60cm then an approximate flow depth of 5m can be determined (Allen 1982). It would be useful to be able to evaluate the minimum time required to produce such a bedform. However this is not simple since it is necessary to consider the lag effects of producing a megaripple from a plane bed, i.e. the delay between a hydrodynamic input, such as an increase in flow velocity and the response of the sedimentary system, such as the formation of megaripples (Allen 1974). Lag (or hysteresis) is a function of the rate of sediment transport (Allen & Friend 1976) but it is difficult to make even rough estimates of this variable. The other measure one needs to take into account is the celerity or bedform migration rate. This may be approximated to the bedform reconstitution time (i.e. the time it takes a bedform to migrate its own wavelength downcurrent), which has been applied to aeolian bedforms by Wilson (1971). Using Wilson's relationship between bedform cross sectional area and time, and taking reasonable limits for the estimate of the sandflow rate, a reconstitution time in terms of tens of hours or days is obtained. Although celerities have been calculated for some modern bedforms, information is lacking for the grain sizes encountered in this example and it is not always clear how applicable celerities obtained in fluvial and tidal environments are to turbidite systems.

Coleman (1969) obtained migration rates of on average 130m/24 hours for bedforms with a waveheight of between 30 and 200cm in the Brahmaputra River; if applied to the Rhinog Formation example this gives a time of approximately 10 hours plus the reaction time, in order to produce the bedform. More information is required to constrain assumptions on the flow and sediment transport characteristics of flows in turbidite basins. However it seems improbable that turbidity currents could maintain high flow velocities for sufficiently long periods of time to produce such large bedforms.

There is evidence that in some cases (e.g. type 1 cross-bedding) the currents which produced the cross-bedding did not only winnow out the fines locally from previously deposited turbidites, but also supplied their own sediment. Some cross-bedded units are particularly rich in volcanic fragments relative to the turbidite sandstones (Crimes 1970a, and see Chapter 7).

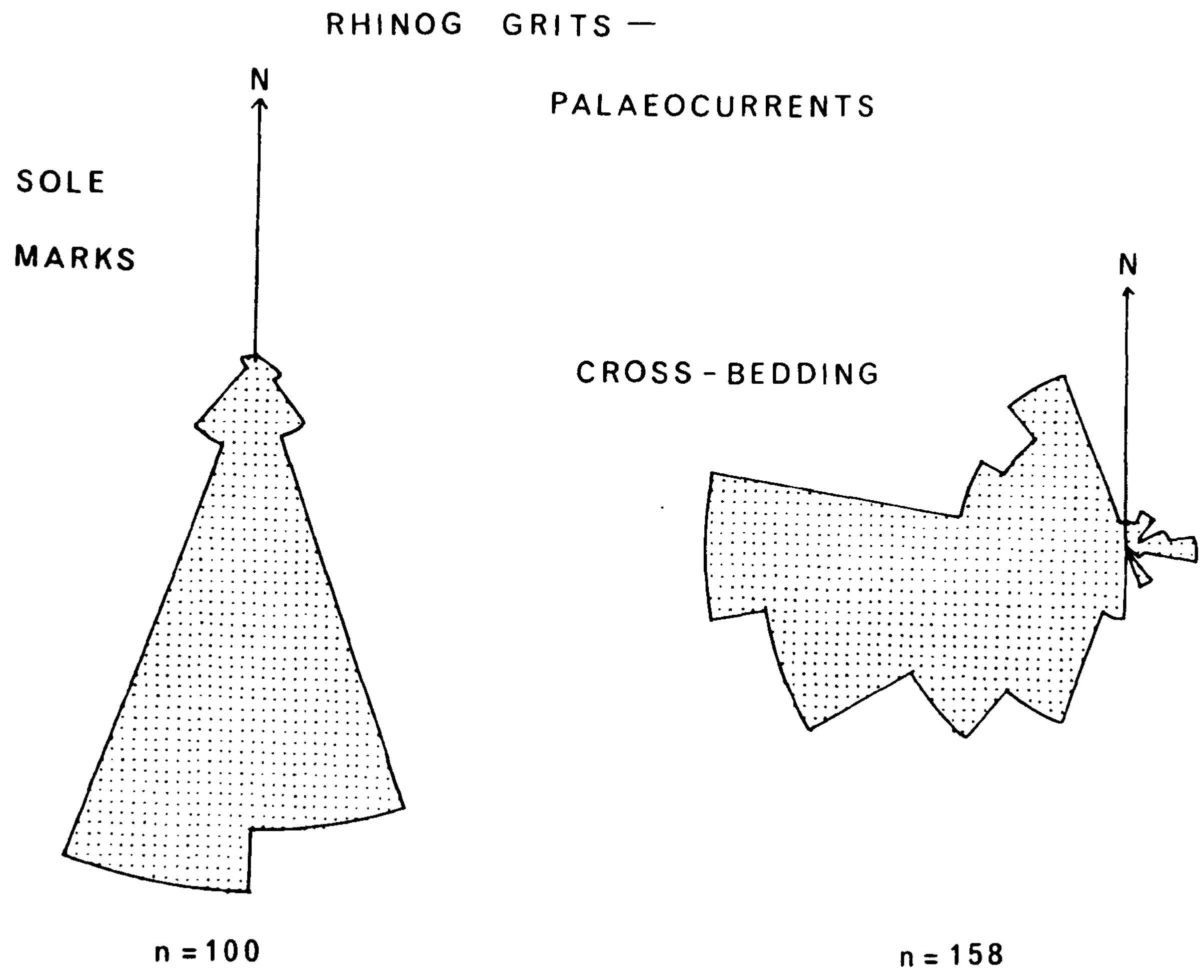
Cross-bedding occurs principally within the Sand-rich Facies, but also occurs in the Thin-bedded, Conglomeratic and Amalgamated Coarse Grained Facies.

Palaeocurrents derived from cross-bedding indicates mainly westerly flow which is approximately perpendicular to the N-S directions derived from sole structures (Crimes 1970a and present data, Fig 4.47). Crimes also plotted the alignment of "washouts" or scours and although these have a variable distribution there was a pronounced maximum in the W-E direction and he suggested that the cross-bedding, the "washouts" and the "massive beds" may have been produced by a similar process.

The cross-bedding is in general produced by relatively dilute flows where traction is the dominant process. However it is important to try to determine the origin of these currents, principally by analogy with modern and ancient examples described in the literature.

Cross-bedding has been described from many turbidite sequences, particularly in sand-rich systems e.g. Dzulynski

FIG 4.47 Palaeocurrents: Rhinog Formation.



et al. (1959): Polish Carpathians in a similar facies of "fluxoturbidites"; Unrug (1963): Polish Carpathians where cross-bedding is associated with beds showing lateral variations in thickness; Craig & Walton (1963): Silurian, southern Scotland where 1m scale cross-bedding indicates flow at variance to some of the sole strucures; Piper (1970): Silurian, western Ireland where cross-bedding is associated with turbidite channels; Chipping (1972): California; Hendry (1972): French Alps where cosets are present; Aalto (1976): Franciscan, California; Cas (1979): Australia, where there is an association of cross-bedding and amalgamated beds; Winn & Dott (1977,1979): Chile; Pickering (1980): Precambrian, Finnmark, Norway; Mutti *et al.* (1981): Hecho basin, Spain; Gokcen & Kelling (1983): Turkey, where cross-bedding occurs on a sandstone-rich low efficiency fan; Helm & Pickering (1985): Precambrian, Jersey.

Some of the cross-bedding described in the above papers are similar to that seen in the Harlech Dome. There are also a number of papers on the Lower Palaeozoic of Quebec and in particular the Cloridorme Formation of Quebec, which describe cross-bedding on a similar scale and with similar characteristics to the Rhinog Formation: e.g. Hubert *et al.* (1970), Rocheleau & Lajoie (1974), Hendry (1978), Walker (1978), Hiscott & Middleton (1979), Johnson & Walker (1979), Strong & Walker (1981), Hein (1982), Hein & Walker (1982) and Pickering & Hiscott (1985).

Cross-bedding may also occur frequently enough to dominate a facies, for instance facies B2 of Mutti (1979), Ricci Lucchi (1984) and facies B2.2 of Pickering *et al.* (1986).

Cross-bedding may occur as a result of three main processes in turbidite basins:

1) Traction at the base of a high density turbidity current.

This process and the nature of the expected deposits is described by Lowe (1982). There are several problems in applying this interpretation to the Rhinog Formation:

a) The sharp tops to most sets of cross-bedding would require near perfect segregation of coarser grains within the flow and deposition of the finer fraction elsewhere (i.e bypassing of sediment).

b) Some sets are overlain only by thin siltstone beds so that there is no evidence for the suspension load of the turbidity current.

c) Cross-bedding is often produced by relatively low density flows, while S1 and S2-type features tend to be produced by much higher density flows.

d) One would not expect development of large bedforms by a turbidity current because of the unsteadiness and short duration of individual flows.

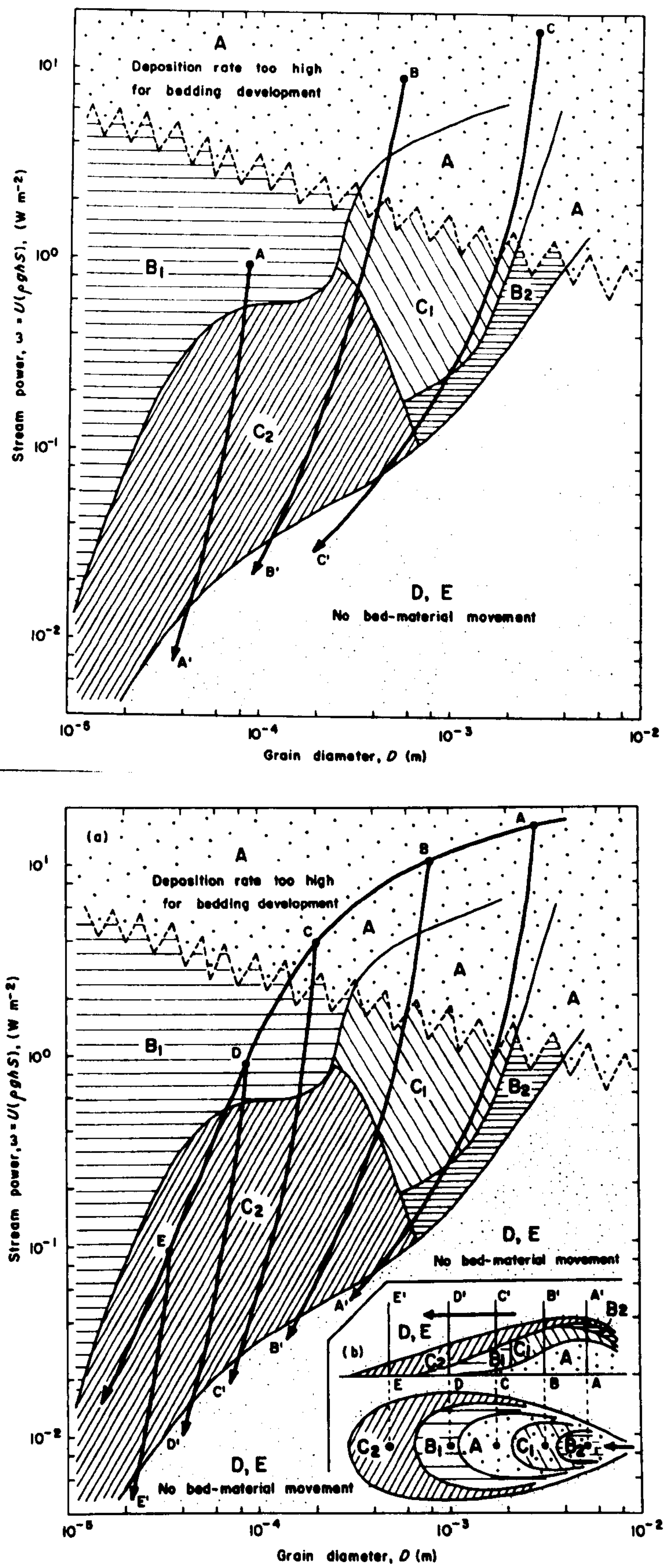
Therefore this seems an unlikely mechanism for the production of most of the cross-bedding. The difference in palaeocurrents between cross-bedding and sole structures is also difficult to explain using the basal traction process. However some of the coarse conglomerate filled scours, the type 7 cross-beds and possibly some of low angle type 6 cross-beds may be produced by this method.

2) Reduction in flow velocity of a turbidity current.

As indicated by Allen (1970), coarser grained turbidites may show a modified Bouma sequence, resulting from waning flow, so that dunes (division C1) occur above parallel lamination (B) and below ripple cross lamination (division C2) (Fig 4.48). Thus grain size is a major control on the type of sequence that tends to occur. However not all coarse grained turbidites show dune cross-bedding largely because turbidity currents tend to decelerate too quickly, i.e they pass too rapidly through the dune field of the bedform stability diagram (Allen 1982, p415) for dunes to

FIG 4.48 Sedimentary structure development in turbidites.

(division C1 is cross-bedded)



(from Allen 1982)

develop and also the flow may be too unsteady.

Although some of the cross-bedding in the Harlech Dome may possibly conform to this model, there are several problems in applying it to the majority:

i) The difference in palaeocurrents between sole structures and cross-bedding. A discrepancy in palaeocurrents between those produced by erosion e.g sole structures and those produced by deposition e.g ripple drift has been recorded in the literature e.g Kelling (1964), Walker (1970). This difference was accounted for by contrasting flow directions in the erosive head with those in the body of the turbidity current so that they need not necessarily flow in parallel directions. However no clear reason is given for large flow divergences.

ii) The base of cross-bedded sets is often sharp and commonly erosive suggesting a clear break in deposition. Similarly the tops of these beds are usually sharp and do not pass upwards into the expected turbidite divisions as predicted by Allen (1970). Instead the bedform is often either amalgamated with the bed above or draped in silt/mud, the latter suggesting a rapid decrease in flow velocity without rapid fallout of sand from suspension.

iii) The cross-bedded unit is usually better sorted and in some cases coarser grained than the bed it overlies. It is possible that winnowing of sediment could occur as a result of waning flow, though it would require a change in flow type from high density (rapidly depositing a wide range of grain sizes) to lower density (winnowing out the fines into suspension). Where the cross-bedded set is coarser grained than the underlying bed then this could be explained by a current which has considerably more energy than the threshold transport velocity of the grains it is carrying. The occurrence of coarser grains would then reflect the patchy distribution of coarse grains being transported as bedload. However one would expect to see a more gradational base to coarser grained units and this hypothesis would also require large gaps in the grain size distribution of the sediment being transported. It is, however, easier to explain the above characteristics by an increase in flow velocity within a surging flow, though whether such an unsteady flow could produce large scale cross-bedding is doubtful.

One way of accounting for the divergence in palaeocurrents is to invoke turbidite deposition derived from the east of a different character to the main northerly derived turbidites in the Rhinog Formation. This would be supported by the scour palaeocurrent data, but such a hypothesis would have similar interpretative problems to those discussed above.

Mutti (1979) described a turbidite channel mouth facies, where bypassing turbidity currents, which were largely non-depositional in this area, reworked sediment and were capable of developing cross-bedding. This zone of bypassing might help explain the lack of mud in the Rhinog Formation but there is little evidence for multi-event channels that could have supplied this zone.

Hein & Walker (1982) provide another model which could account for cross-bedding in turbidites. The Cap Enrage Formation, Quebec, Canada, which they describe, has cross-bedding very similar to the Harlech Dome (e.g sets about 45-50cm thick with multiple sets, an association with pebbly sandstones and "massive" sandstones). They interpret the environmental setting of these rocks as comprising of braided turbidite channels flanked by terraces, the cross-bedding being produced by overspill from the channels onto the terrace. This is used to account for the variability in palaeocurrents obtained from the cross-bedding. This model may be applicable in part to the Rhinog Formation, though it lacks the deep, incisive braided channels of the Cap Enrage Formation.

Winn & Dott (1977) interpret large scale (2-4m amplitude) dunes as the result of low density turbidity currents reworking by traction the deposits of high density turbidity currents, a mechanism also described by Lowe (1982). However low density flows are usually also characterised by lower flow velocities and therefore their ability to transport coarse sediment must be questioned. Since they predict that these bedforms would require hours or days to form they regard these as composite in structure, the result of deposition from several events. Although this model may be used to interpret the coset bed type in the Rhinog Formation, the lack of mud drapes between sets and lack of erosion in some cases between sets argues against deposition by several turbidity current events.

Other possible ways of accounting for divergences in palaeocurrents might be reflecting or contained turbidity currents produced by "sloshing" of flows in a confined basin (Pickering & Hiscott 1985), though evidence for substantial decreases in flow velocity would be expected between flows. Another possibility is the production of post-turbidity current "residual" flows, the characteristics of which are not well understood.

Modern analogues for these megaripples are rare. Sediment waves or megaripples, interpreted as being produced by low density turbidity currents are recorded by Normark *et al.* (1980) at abyssal depths on the Monterey Fan. Bouma & Treadwell (1983) recorded three types of dune form on the lower Colombian continental rise, but in general these bedforms have a much lower amplitude to wavelength ratio and are much larger scale features (having amplitudes of 3-100m, and wavelengths of 0.5-6km) compared to those of the Rhinog Formation. Studies of smaller dunes are however limited by the resolution of the seismic records.

3) Reworking by constricted bottom currents.

In certain circumstances bottom currents can be constricted and attain the competence to transport sand grade sediment (Stow 1982) and may have the persistence to produce bedforms. These flows are usually dilute and thus traction processes are dominant. However most of the cross-bedded sets in the Harlech Dome are much coarser grained than most described contourites (Bouma 1972, Stow & Lovell 1979). Geostrophic currents, either driven by an undercurrent of cold dense water (e.g. the Labrador Current) or by warm saline water (e.g. the Mediterranean Outflow Current) may produce flow velocities of 2-20cm/sec, though exceptionally flow velocities may reach up to 300cm/sec where constricted (Stow 1982, Stow & Holbrook 1984). Thus constricted flows may, for a short time, reach the flow velocities necessary to transport coarse sediment. Geostrophic currents control the distribution of large sediment drifts e.g the Faro Drift- 50km long and 300m thick and these may show surface evidence for the action of strong bottom currents (Faugeres *et al.* 1984). Most described geostrophic currents have been from oceanic settings where large water density contrasts exist to drive the geostrophic currents which often flow at abyssal depths. However Johnson *et al.* (1980) describe sedimentologic evidence for wind-driven geostrophic currents at 200m depth in Lake Superior, which may intensify in strength during periods of storm activity. Whether the Cambrian Welsh Basin could have had similar currents is not clear.

Other possible currents active in deep basins include canyon currents which may travel up or down canyon, and show alternations in direction related to tidal periodicity as well as the affects of internal waves. However flow velocities rarely exceed 50cm/sec, though this is usually sufficient to transport large quantities of fine sediment, usually downcanyon (Shepard & Marshall 1978). Valentine *et al.* (1984) record dunes with wave heights of 0.3-3m and wavelengths of up to 15m in Oceanographer Canyon off the Atlantic coast of the USA. The dunes are composed of coarse sand, which they calculate to require velocities of at least 70cm/sec to transport (thus they are of comparable grain size and dimensions to the bedforms of the Harlech Dome). These bottom currents also winnow out finer sand. However although the internal structures of these bedforms are not known, bottom current velocity measurements indicate that the bedforms probably show foreset geometries indicative of reversing flow, which is not seen in the bedforms from the Rhinog Formation. There is also no direct evidence that the

Rhinog Formation shows the characteristics of canyon fill.

In summary, cross-bedding in the Rhinog and Barmouth Formations indicates mainly flow from the east, whereas the majority of sole structures indicate that the turbidite sandstones were derived from the north (in the Rhinog Formation) and south (in the Barmouth Formation). These flows were dilute and some deposited sediment that was coarser and richer in volcanic clasts than the beds below (see chapter 7). Some cross-bed sets are relatively thick (e.g. 60cm) and were probably deposited on a time scale of the order of tens of hours, so were produced by relatively sustained flows.

Since some cross-bedding forms discrete beds and palaeocurrents indicate flow from the east this suggests that it was not deposited from normal turbidites. However no known geostrophic or canyon currents capable of forming this type of cross-bedding has been found. Therefore a turbidite origin is favoured for the cross-bedding, though these flows must have been traction dominated and have flowed from the east. Whether turbidity currents are capable of maintaining the flow velocities necessary to produce such large bedforms is unclear.

The flows which deposited the cross-bedding could have:

a) Originated on a slope with a volcanic-rich supply to the east, or

b) Formed by westward flowing crevasse splay from a northerly derived turbidite channel which lay to the east of the present Harlech Dome outcrop and had a coarser, more volcanic-rich source.

4.7 : Lateral Variability.

Analysis of the lateral variability of beds and packets is important in order to determine the geometry of units of different hierarchies in the Harlech Dome turbidite system. An understanding of unit geometries then aids interpretation of the nature of the turbidite system. Lateral variability was investigated at several different scales:

1) Outcrop Scale. Lateral control of the order of 10m.

In general most beds and virtually all packets are tabular when examined at this scale, usually with parallel upper and lower boundaries overall, though irregular scours may be superimposed. When followed laterally most beds show little change in thickness apart from beds in the Amalgamated Coarse Grained Facies where considerable lateral changes may occur and amalgamated bed contacts are diffuse and irregular.

Scours are common in the Rhinog and Barmouth Formations; most appear to be symmetrical and relatively shallow (usually with an amplitude of less than 1m) and they do not normally show any consistent direction of downcutting. Therefore scours in general seem to represent localised areas of erosion at the base of a sheet flow rather than a relatively narrowly confined channel. However certain scours have a coarse grained and relatively sharp topped fill or lag. Although scours may occur preferentially at certain horizons, the individual scour-and-fill units are highly lensoid even on this scale. Some shallow pebble-lined fills show evidence for successive scour-and-fill events (e.g Bwlch y Llan [SH 621 178]). Certain structures may indicate interdigitation between successive turbidite events (Fig 4.49, Foel Wen [SH 628 265]). The interdigitation could have been produced in two ways:

i) by abrupt depositional thinning of the lower bed

FIG 4.49 Lateral variation in turbidite beds, Rhinog Formation, Foel Wen.

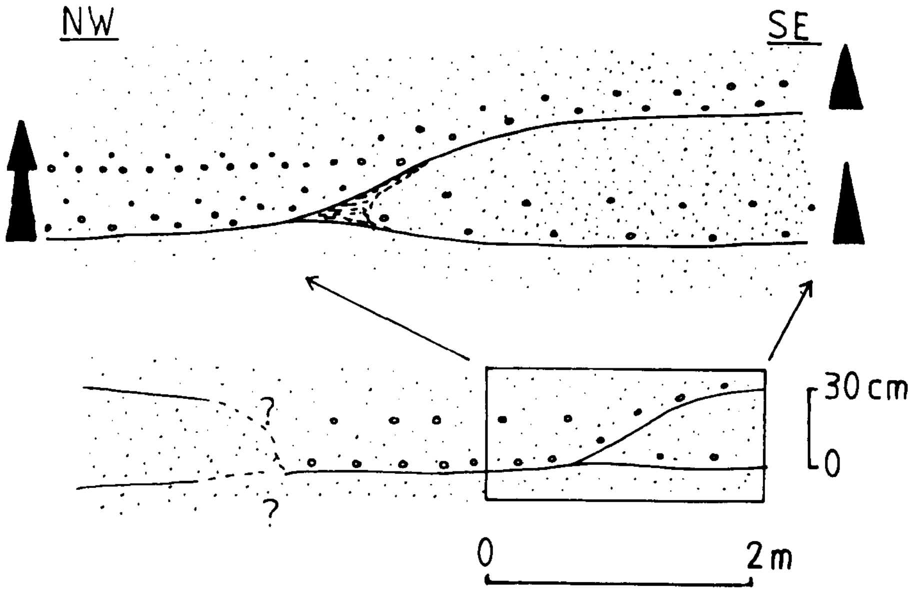
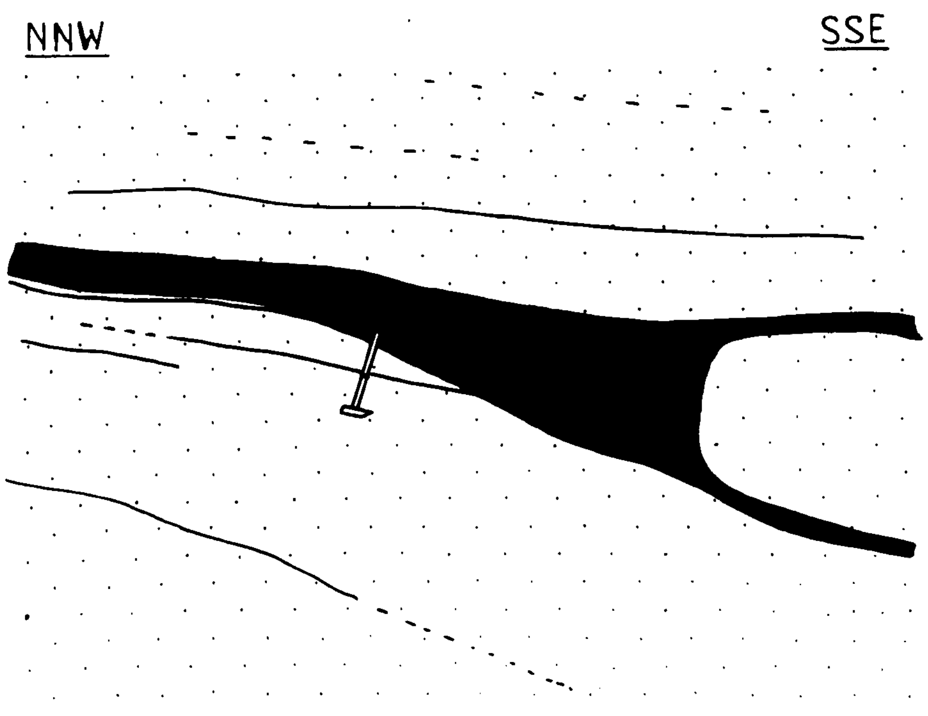


FIG 4.50 Lateral variation in turbidite beds, Barmouth Formation, Y Garn.



and then infill of the topography by the succeeding turbidite sandstone (which shows multiple grading).

ii) by scouring down through the lower bed, which was originally of constant thickness (on the scale of the outcrop), down to the underlying bed's lower boundary). The scour was then infilled by the upper bed. The latter hypothesis appears to be more likely.

At Y Garn [SH 702 230] a thick relatively massive sandstone bed, when traced from south-south-east to north-north-west thins to nothing with a nearly vertical boundary (Fig 4.50). In the other direction the sandstone bed is laterally persistent and of constant thickness. The sandstone bed is underlain and overlain by cleaved siltstones. This exposure therefore shows very rapid lateral changes, possibly indicating that sand deposition was confined within a shallow but steep sided scour.

Some sandstone beds in the F Rock area (Section 2, Fig 4.40) show irregular tops which appear to have been eroded and infilled with Thin Bedded Facies. This may account for many of the lateral differences in bed thickness which occur on the outcrop scale.

Other types of lateral variation may also occur on the outcrop scale especially in the Amalgamated Coarse Grained Facies e.g grain size, grading.

2) <u>Large Outcrop Scale.</u> Lateral control of the order of tens of metres to a maximum of 175m.

The geometry of beds was investigated at the large outcrop scale in various localities. In particular an area of good three dimensional exposure of Rhinog Formation was examined around F Rock. This locality is north of the Roman Steps and east of Cwm Bychan (Fig 4.40) [SH 666 312]. Several sections are described:

<u>Section 1.</u>

Overall sandstone packets have a tabular geometry on the scale of the outcrop (55m long), though individual beds vary considerably laterally. This sequence can be divided into 6 units (Figs 4.51, 4.52):

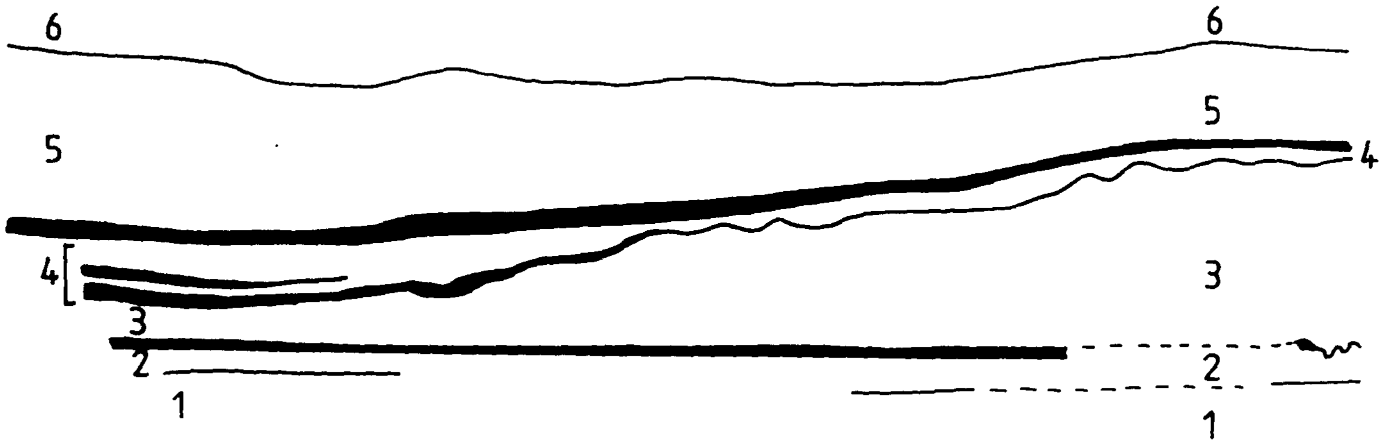
1) T_a turbidite sandstones. The top of the uppermost bed is locally cross-bedded with a set height of up to 10cm and indicates flow towards the southwest.

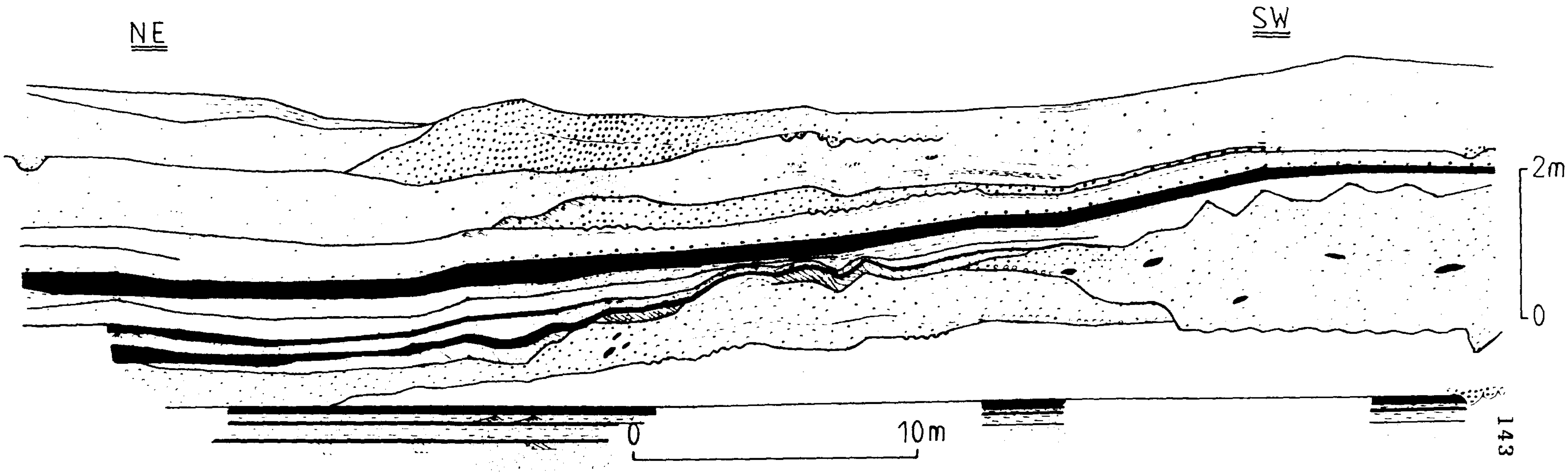
2) Thin bedded unit. Two parallel laminated (T_b) fine sandstones, one of which passes laterally into cross laminated (T_c) sandstone with ripple drift. The lower bed is 19-24cm thick, the upper bed 10-13cm. Thus over the width of the outcrop they are sheet-like with planar bases and tops, though changes occur laterally in their internal structures. These sandstone beds locally grade up into siltstones.

3) Planar, horizontally-based unit with a relatively planar, inclined top. This unit is thickest to the southwest and thins towards the northeast. If thickness measurements are taken at 10m intervals from southwest to northeast then the thickness of the unit changes thus: 283cm, 228cm, 175cm, 104cm, 55cm. Much of the decrease in thickness is related to the medium sandstone bed at the base of the unit which ranges in thickness from 0-113cm. This sandstone bed is overlain by highly variable beds which are in general coarser grained and include abundant evidence for traction (e.g low angle cross stratification, Figs 4.53, 4.54). Overall unit 3 coarsens upwards and also towards the southwest. The bases of some of these beds are commonly erosive and loaded, are often filled with the coarsest grains (mainly granules) and may have abrupt or graded tops (Fig 4.55).

Intraclasts in unit 3 are often angular (Fig 4.56) and are concentrated in certain parts of the section, indicating erosion nearby, perhaps by localised vortices within the turbidity current concentrating scour at certain points. The largest and most angular intraclasts tend to occur nearest to the region of erosion (i.e. the site of intraclast production). Downcurrent from this point, buoyancy effects would be expected to become more important and intraclasts become smaller, more rounded and tend to occur nearer the

FIGS 4.51 and 4.52   Compensation Cycle, Rhinog Formation, F Rock section 1.





Sedimentary structures, Rhinog Formation, F Rock section 1.

FIG 4.53

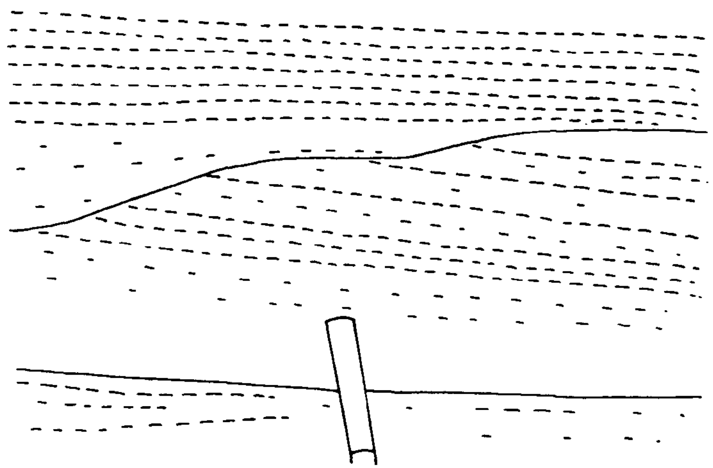


FIG 4.54

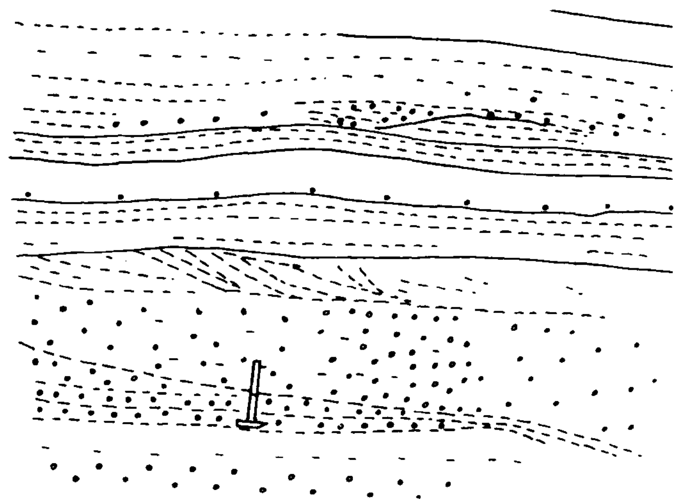


FIG 4.55

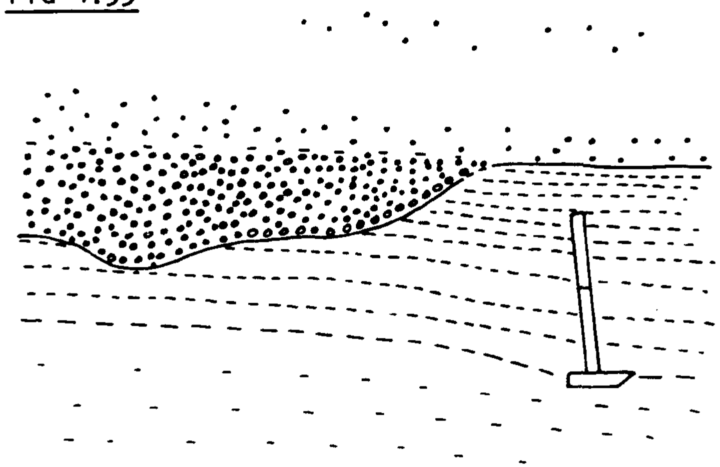
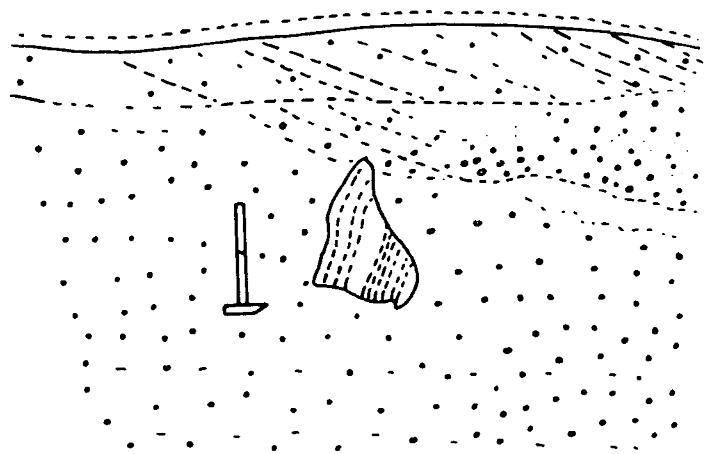


FIG 4.56



tops of beds (Mutti & Nilsen 1981). Most of the intraclasts in this section seem to have been produced near to source, though their concentration at a particular horizon may be difficult to distinguish from a load-affected fine grained bed. However a lack of soft-sedimentary down-loaded laminae and the significant amount of rotation of some of the blocks argues for depositionally produced intraclasts.

Erosive bed contacts in this unit e.g. at the base of the amalgamated coarse grained bed in the far southwest of the section, may indicate channelisation. However although this bed thins markedly towards the northeast, as a result of a relatively steep scour surface, lateral changes in grain size are generally more gradational. The scour was probably cut and filled in a single flow event.

The base of the unit is largely horizontal so most of the overall thickness changes are the result of relief of the order of 1m on the upper surface of the unit. The upper boundary of this unit is probably, at least in part, depositional in origin since in general the grain size of the unit gets coarser grained from northeast to southwest. The grain size variation may result from slightly higher flow velocities or an increase in the supply of coarse sediment at the same velocity (due to patchiness of the distribution of coarse sediment within the flow), in the southwestern part of the section. The other possible way this geometry may be produced is by successive erosion of initially tabular units and then infill of these scours. Although this may explain the large scour in the far southwest of the section, in the northeastern part of the section the dipping contact between the lower medium grained sandstone bed and the coarse grained sandstone above parallels the topography on the upper surface of the unit.

The upper boundary is thus mainly non-erosive. The general northeasterly dip of the upper boundary of unit 3 is probably not due to scour but to differential deposition, possibly on a small aggrading lobe rather than channel setting.

The topography on the upper surface of the unit was however modified slightly by traction currents (section 4.6). These currents produced cross-bedding in slightly better sorted sandstone than that which they overlie. Set thicknesses range up to 25cm. The cross-bedded sets truncate some of the laminations in the beds below (mainly in the southwestern part of the section), indicating reworking of the underlying bed, i.e. mainly winnowing with only minor erosion. The cross-bedding was produced by currents flowing towards the southwest, which was probably at an oblique angle to the dominant currents which deposited this sequence. Scours tentatively indicate flow into the plane of the section (i.e. scours are aligned approximately N-S).

4) A unit composed of three T_{bde} turbidite beds, which grade upwards from fine sandstone to very fine sandstone. These beds tend to thicken into the depression in the upper boundary of unit 3 (i.e. towards the northeast) and thin towards its crest (i.e towards the southwest). For instance one bed thickens in a traverse from southwest to northeast (thickness measurements taken at 10m intervals): 0, 4cm, 6cm, 12cm, 15cm. The bed which shows the largest variation in thickness is the lowest bed; successive beds reduced the topographic relief as preferential deposition occurred in the depression and there is progressive onlap onto the upper boundary of unit 3. Thus successive small-scale turbidity currents were influenced by the pre-existing seabed topography, which to some extent controlled their deposition.

5) A more tabular unit which thins slightly towards the southwest, infilling the last remnants of relief left unfilled by unit 4. The unit is dominated by sand-rich T_a turbidites, some of which show good grading. Bouma sequences present include T_{ab} and T_{abc}, the latter sometimes showing convolute lamination. There is abundant evidence for traction, including both scour and fill structures with abrupt or graded tops, commonly filled with coarser sandstone than the surrounding sediment. Low angle cross-bedding is also common.

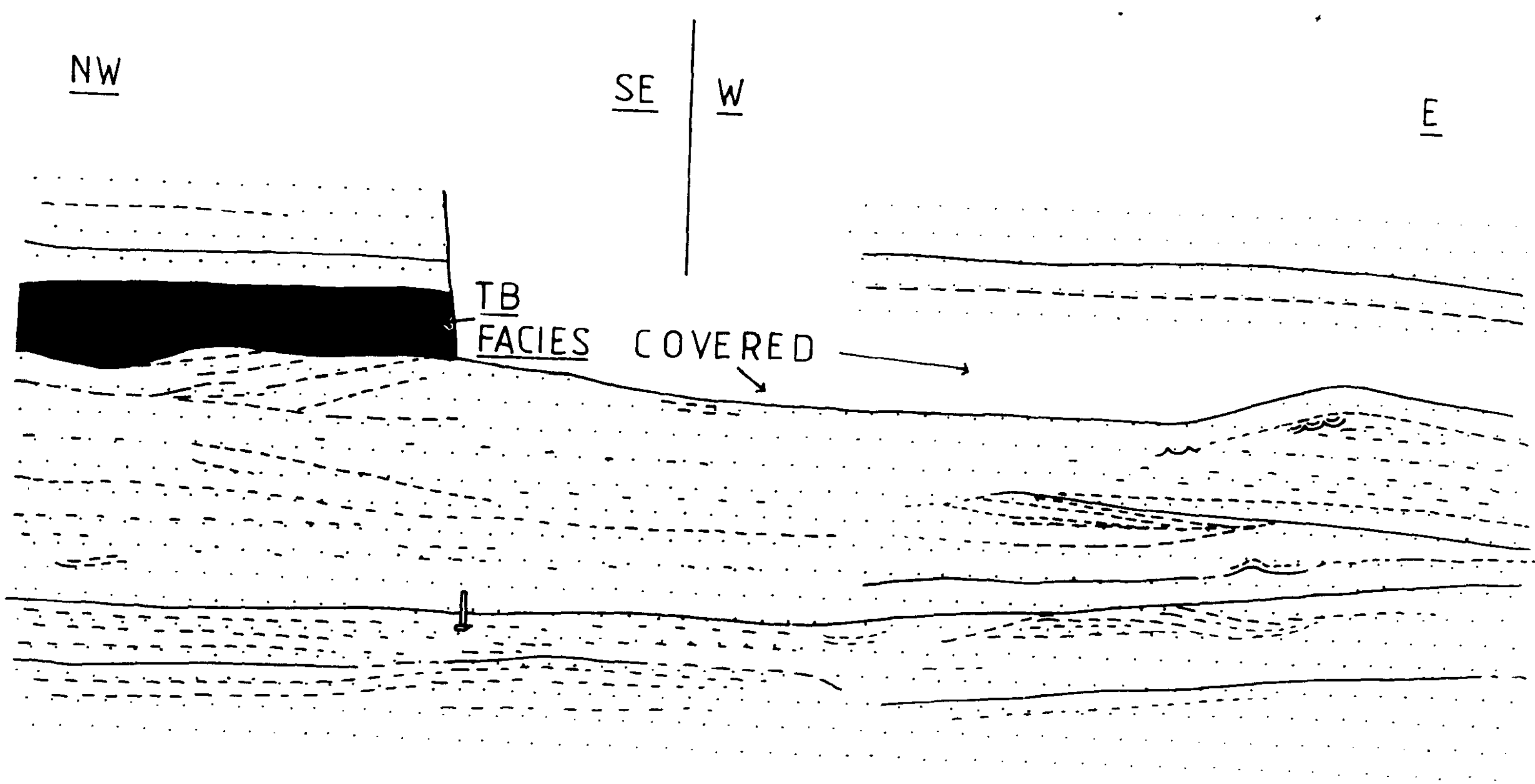
6) Thin bedded facies- T_{cde}, T_{be} turbidites.

The relationship between units 3 and 4 are similar to what Mutti & Sonnino (1981) refer to as a Compensation Cycle. These authors describe features resulting from progressive smoothing of depositional relief which they regard as being characteristic of fan lobe aggradation. They suggest that turbidites will tend to smooth out local topography, producing metre-scale thickening upward sequences in the depressions and thinning upward sequences over the crests. This locality seems to show sequences of a similar type and scale to those recorded by Mutti & Sonnino. Such sequences are likely to occur perpendicular to the dominant flow direction, whereas sections parallel to the flow direction tend to show much less lateral variability.
If we assume that flow in the F Rock section was predominantly from the north (using tentative scour evidence and palaeocurrent data from elsewhere), then the section described is oblique to the major north to south flow direction, thus accounting for the bed thickness variability. This interpretation may also explain the parallel nature of beds at many outcrops in the Harlech Dome (?parallel to flow) relative to others (?oblique to flow).
Topography produced by differential sandstone deposition was sufficient to control deposition of later turbidites (thinner bedded turbidites more than thicker bedded turbidites), resulting in preferential deposition in the depressions. This "compensating" effect in the sedimentary system means that despite differences in individual bed thickness there is a tendency for sand packets to be tabular. A combination of scours and compensation cycles suggest a close interplay between shallow, short-lived channel-type processes (as seen in the far south-west of the section) and aggradational lobe-type processes (between units 3 and 4).

<u>Section 2.</u>

This section (Fig 4.57) shows a sequence of both parallel and low angle trough cross-bedded sandstones; the

FIG 4.57 Section 2, Rhinog Formation, F Rock.



latter has a maximum amplitude of 20cm and tentatively indicates flow towards the south-south-east. Beds clearly thicken and thin in the plane of the section. The upper boundary of the sandstone unit has up to 50cm relief and this is infilled with Thin Bedded Facies. At the northwest part of the section relief on the upper boundary is the result of a megaripple formset. Thus there is strong evidence here for traction reworking of sediment producing beds that are highly variable laterally.

Section 3.

This section (Fig 4.58) includes the thick (up to 60cm thick) megaripple cross-bedded granule conglomerate bed described in Section 4.6 (cross-bedding type 1). This bed is very variable in thickness, lensing out between points C and D and can also be followed in section 3. The beds below are very variable in thickness and grain size partly due to lensing and loading disturbance. However the beds above the cross-bedded set are, in contrast, relatively constant in thickness. The thick massive bed (c. 2.5m thick), for instance, can be followed throughout much of the F Rock area and in general has a non-erosive, planar base.

Section 4.

This section (Fig 4.59) shows considerable lateral variation in bed thickness from west to east. It can be divided into four units:

1) Unit 1 is mainly comprised of massive beds of medium sandstone up to 183cm thick. These beds commonly have erosive, loaded bases associated with injection of finer grained sediment. Bouma T_{ab} sequences are most common, though the parallel laminated part only occupies a small part of the total bed thickness. Internal scours with graded coarse grained fills are common and are up to 20cm deep.

Coarse sandstone bed "A" thickens and thins rapidly from a maximum of 98cm to 67cm at the western end of the section. This bed downcuts into the bed below, so that in the central part of the section the bed below is absent.

FIG 4.58 Section 3, Rhinog Formation, F Rock.

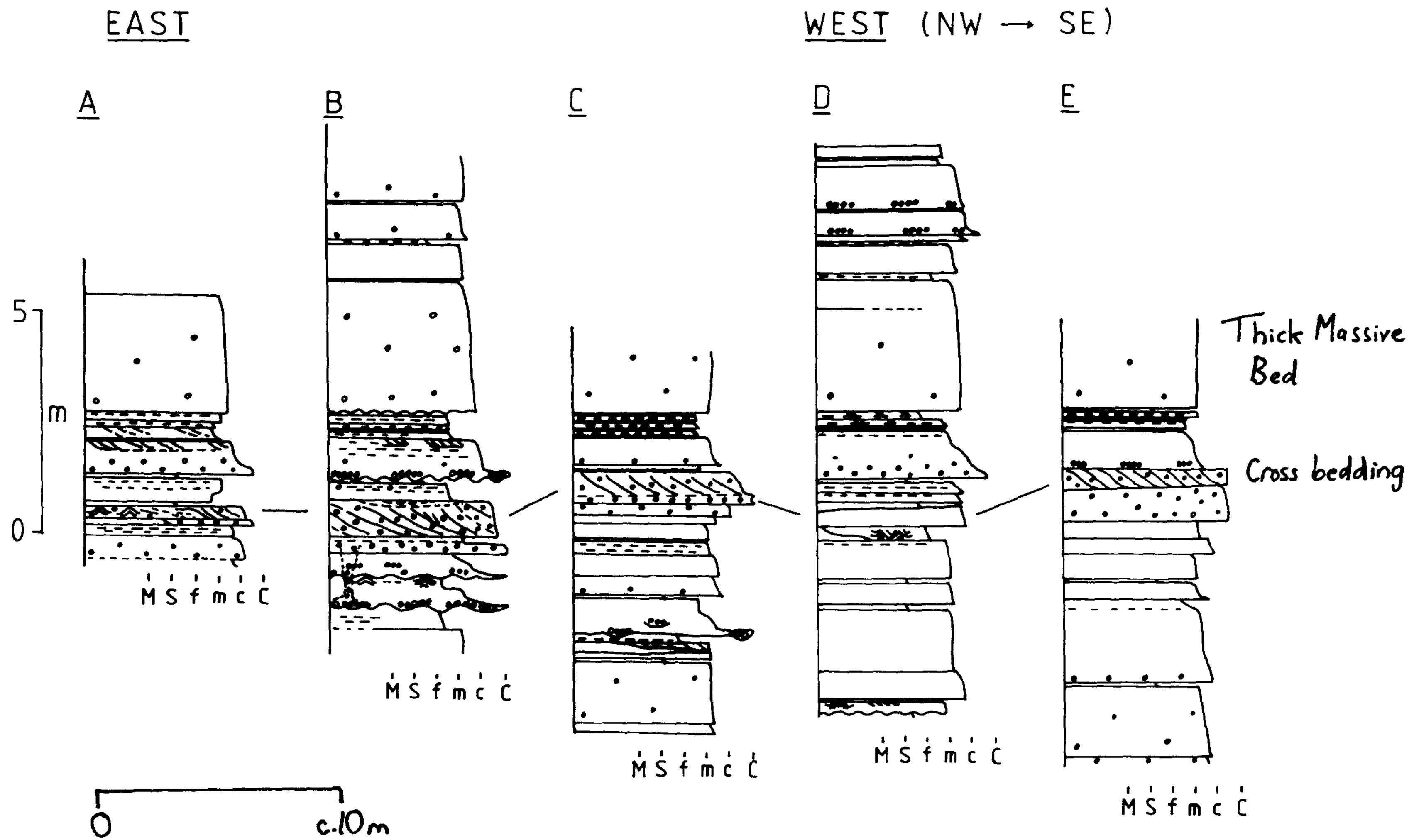
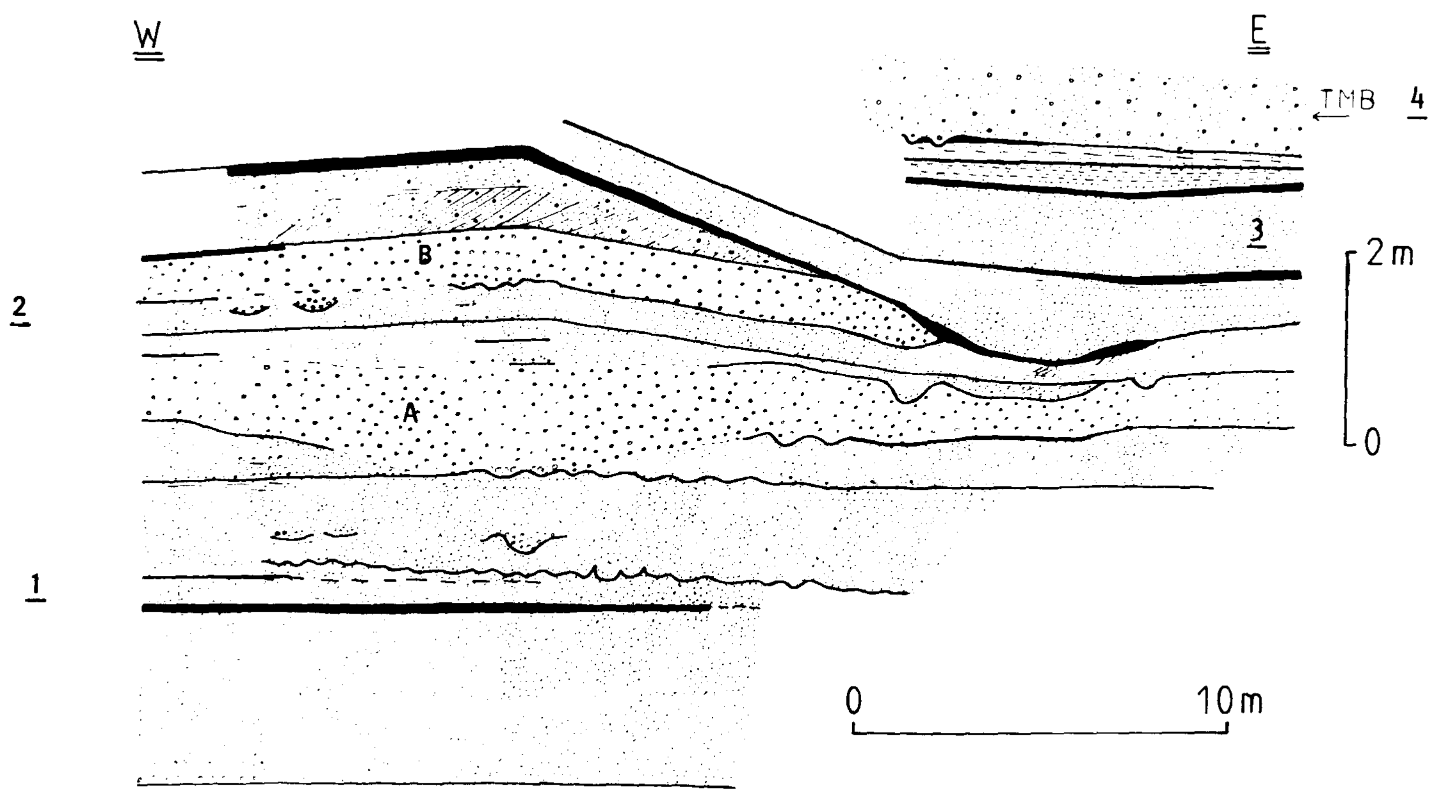


FIG 4.59 Section 4, Rhinog Formation, F Rock.



These features indicate scour and fill processes, with possible short-lived channelised flow of coarse sand grains in the scours. The presence of repeated grading (usually coarse-tail), scour and fill (S1) structures together with well developed inverse grading indicate deposition from high density turbidity currents (Lowe 1982).

2) Unit 2 is similar to unit 1 in its lower part. However individual beds thin markedly towards the east. Coarse sandstone bed "B" thins from 41cm to 0 laterally (west to east) within 1m. The overlying set of cross-bedding indicates flow towards the west and also thins rapidly towards the east. The upper surface of the unit truncates the bed boundaries and is responsible for the bed thickness differences below it. Therefore the upper bed boundary is erosive, the erosion probably post-dating the deposition of the youngest bed of unit 2, though it is possible that the erosion pre-dated the cross-bedding, which was produced by traction current reworking.

3) Unit 3 is comprised of thick (17-81cm) sandstone beds- largely Bouma T_a and T_b. The lower bed drapes relief on the upper boundary of unit 2 and appears to remain relatively constant in thickness, though what the overlying beds do laterally is unclear. It is probable, however that the erosional hollow influenced sedimentation of these overlying beds.

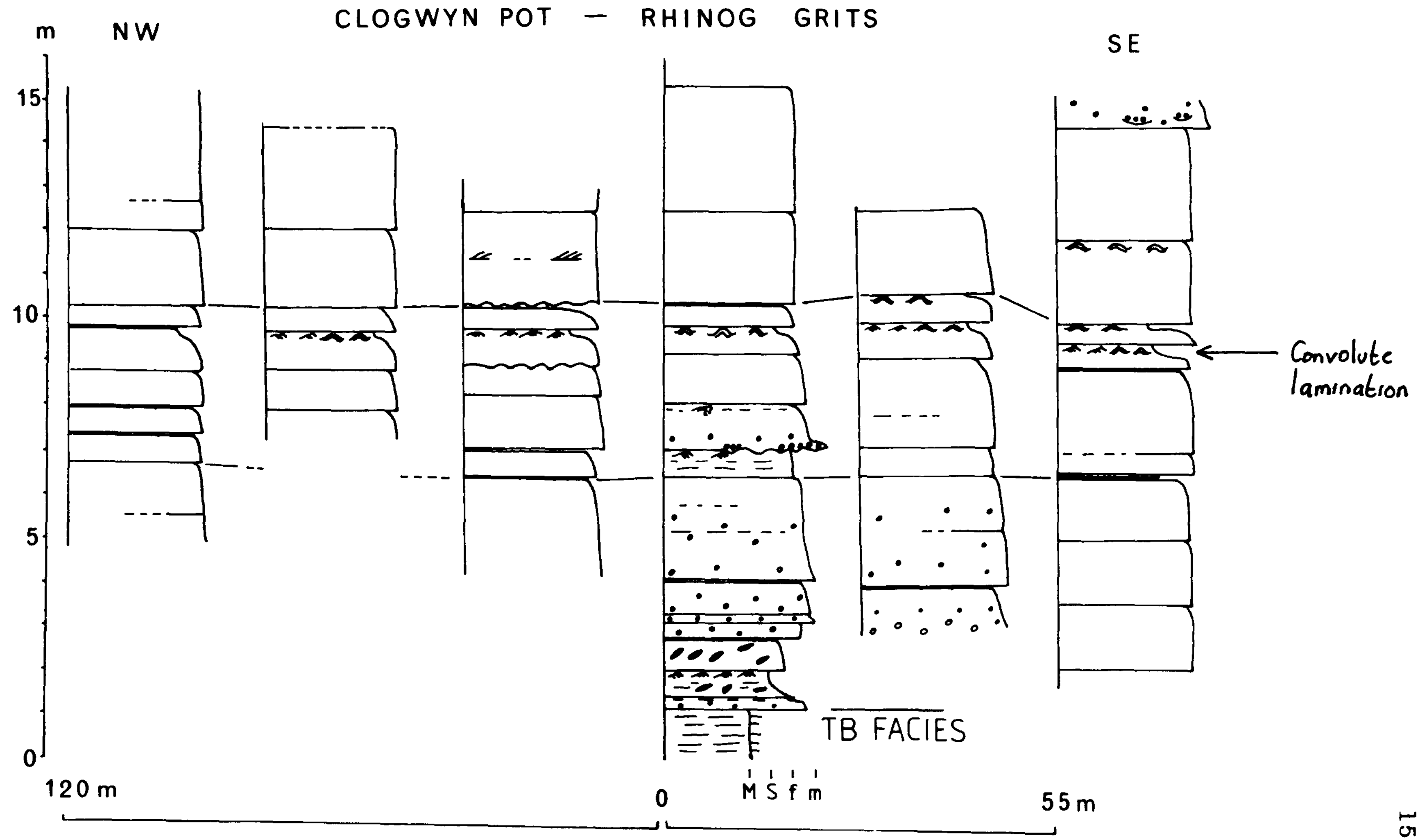
4) Unit 4 is a very thick massive bed (TMB in Fig 4.59) with occasional dispersed granule grains. This thick massive bed occurs in sections 3 and 4.

Two other large outcrops were looked at in order to determine lateral changes in bed characteristics:

Clogwyn Pot- (1km north of the Roman Steps [SH 658 308]).

This section (Fig 4.60) shows that overall the sandstone packets tend to be tabular, though within these packets some beds may vary laterally in bed characteristics. For instance a graded bed with convolute laminated top

FIG 4.60



CLOGWYN POT — RHINOG GRITS
m
NW
SE
15
10
5
0
Convolute lamination
TB FACIES
M S f m
120 m
0
55 m

(T_{ac}) passes laterally into only a graded division (T_a). Lateral variations in amalgamated beds are most pronounced. Some beds, however remain relatively constant in thickness laterally. The section is roughly parallel to the main flow direction and may therefore explain the relative continuity along section. There do not appear to be any systematic downcurrent trends.

Llyn Du: East and West logs- (c.750m south of the Roman Steps [SH 655 293]).

The two logs are approximately 150m apart.

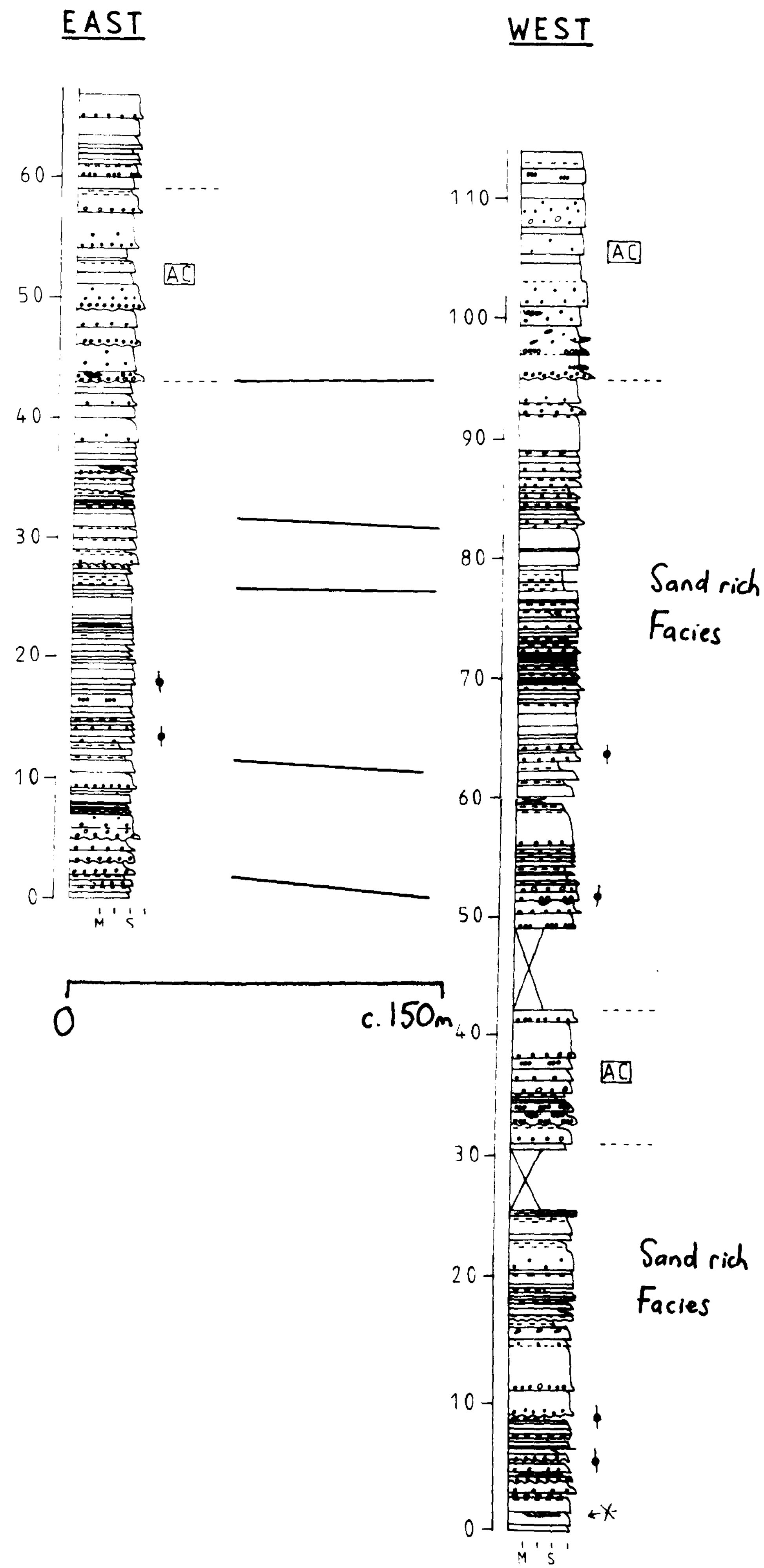
Comparison of the two logs (Fig 4.61) shows that there is considerable lateral variation in bed thickness especially in the Amalgamated Coarse grained Facies. For instance isolated scour and fill structures in the West log become slightly more continuous as part of the scoured and graded lower portion of the bed in the East log; so there tends to be increasing amalgamation towards the west. Other lateral changes also occur: T_{abc} bed at 36m on East log passes laterally into T_{ab} on West log; a T_{ab} bed at 57m on East log passes laterally into T_a at 108m on West log.

Therefore lateral changes occur in bed thickness, in the nature of bases of beds e.g. from planar to irregular, in grain size- especially at the base of beds, in the nature and degree of grading and in sedimentary stuctures.

Most of the lateral changes in bed thickness are the result of differences in the degree of amalgamation. For instance the series of beds (each c. 50cm thick) at 30m on East log; its lateral equivalents at 80m on West log is of similar overall thickness but is generally thicker bedded. This implies an increase in the amount of amalgamation for this particular part of the sequence from east to west. Determining the products of single turbidity current events in such sequences is very difficult to resolve and this makes interpretation of vertical sequences and distinguishing between successive turbidity currents and the products of a single surging turbidity current

FIG 4.61

LLYN DU



problematical.

Lateral changes are not consistent, for as well as the increase in amalgamation from east to west described above for part of the sequence, the West log is also thinner bedded around 70m relative to its lateral equivalent around 20m on East log. Thus lateral changes are not simple or consistent in any particular direction.

At this scale the same general trends can be located on each of the logs. Although individual beds may vary laterally, in general packets of similar beds or amalgamated beds are relatively tabular.

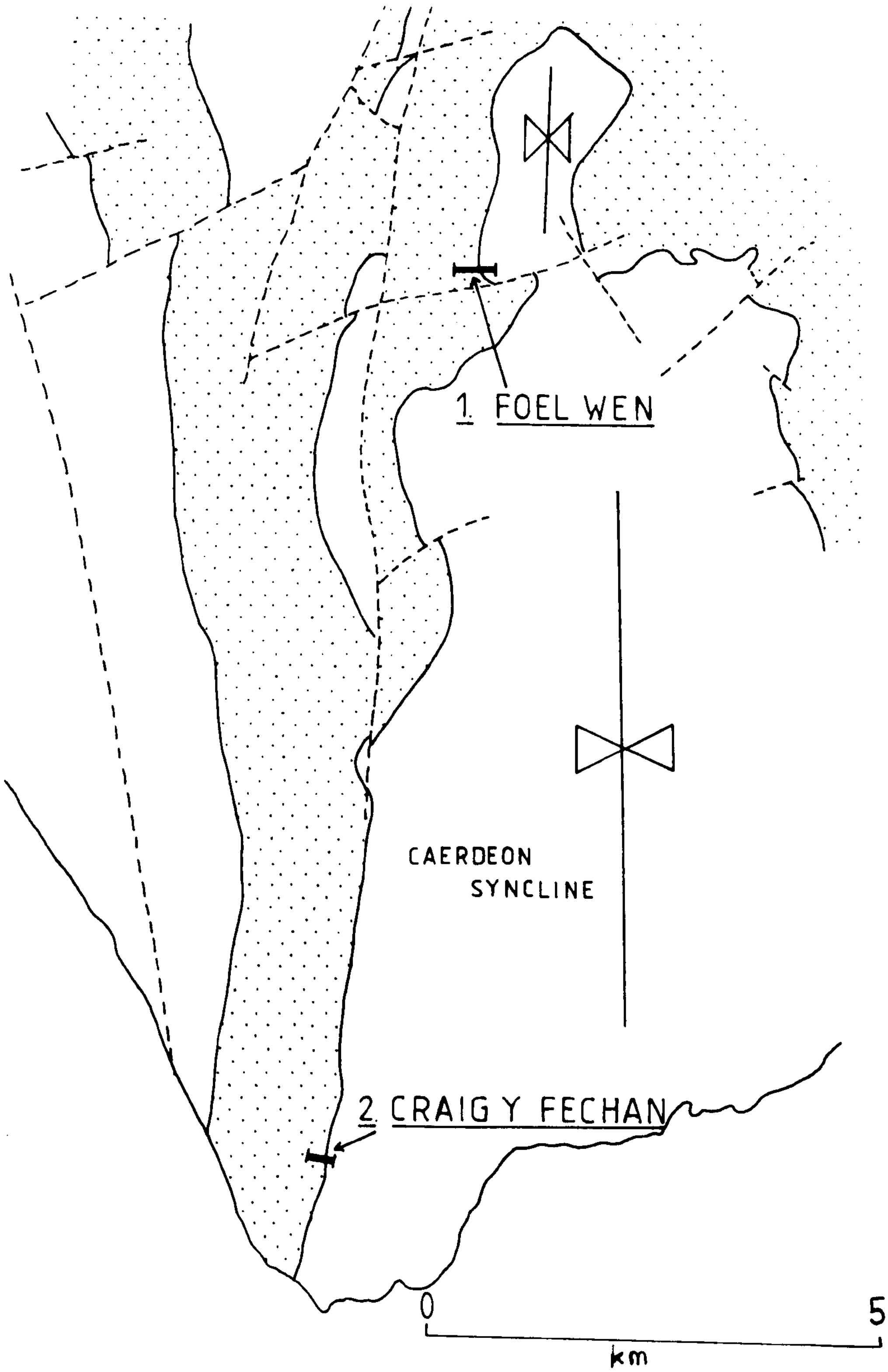
3) Large Scale. Lateral distances of the order of kilometres.

Comparison of sequences laterally on the kilometre scale is difficult due to faulting and a lack of good marker horizons within the Rhinog Formation. Thus it was necessary to compare sequences at horizons where the stratigraphic position was accurately known, at levels interpreted to be broadly synchronous:

i) Top of the Rhinog Formation. (Figs 4.62, 4.63)

Foel Wen in Cwm Nantcol [SH 628 265] is 10km north of Craig y Fechan [SH 615 173] near Barmouth. In general the Foel Wen section is coarser grained with more frequent erosive bases and coarse grained scour and fill structures. Amalgamation is slightly more common and bed contacts are more diffuse and irregular than the Craig y Fechan section. From Foel Wen (100-140m) to Craig y Fechan (50-80m) there is a lateral change from Amalgamated Coarse Grained Facies to Sand-rich Facies. There also appears to be a decrease in thickness of this part of the sequence from north to south, which is probably the result of continued coarse turbidite deposition in the Foel Wen region since the Thin Bedded Facies below is of approximately equal thickness. Thus there does not appear to be a greater amount of downcutting in the

FIG 4.62 Distribution of Top of Rhinog Formation logs.

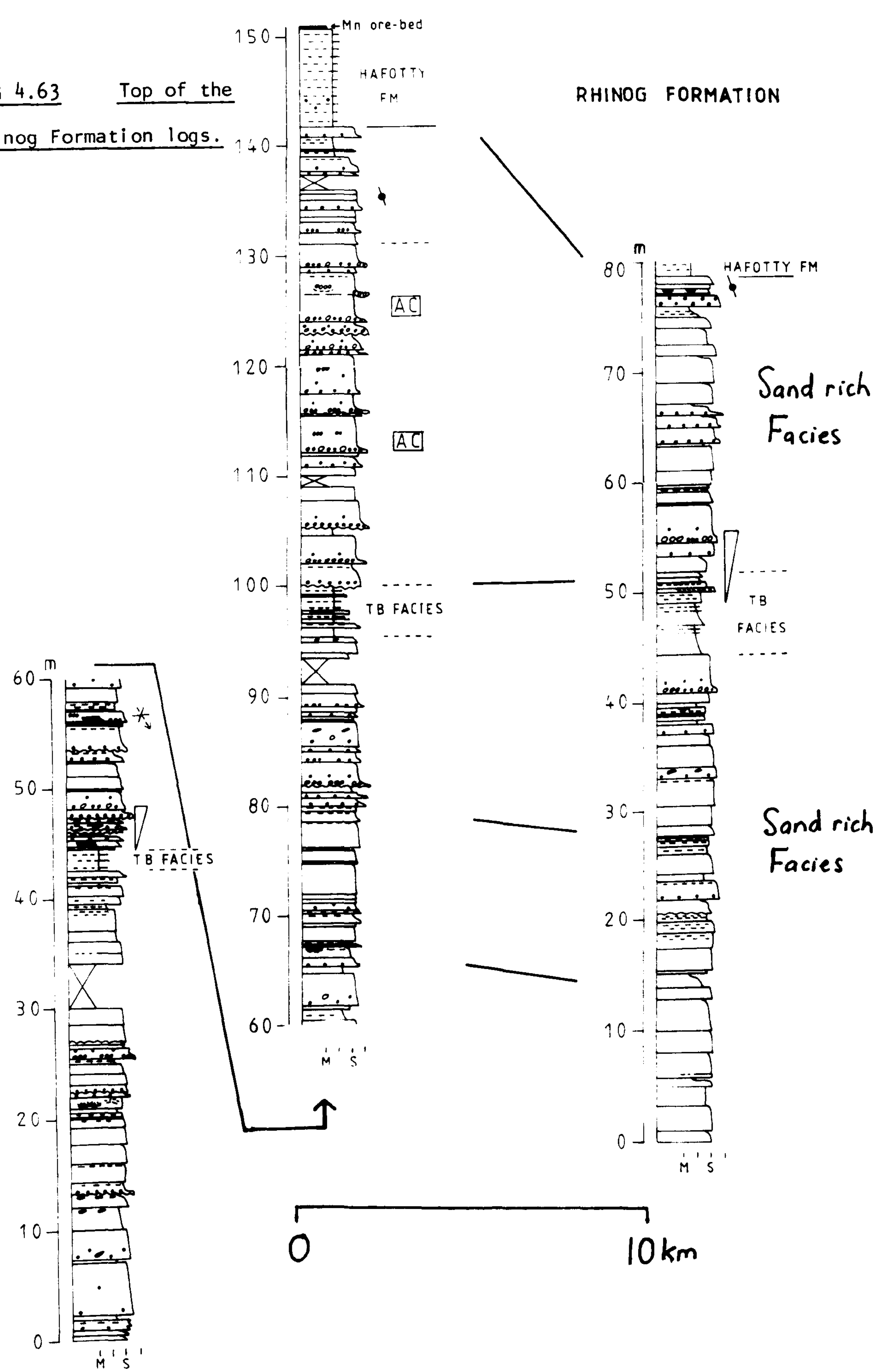


FOEL WEN

CRAIG Y FECHAN

FIG 4.63 Top of the Rhinog Formation logs.

RHINOG FORMATION



Foel Wen section at this horizon relative to the Craig y Fechan section but the top of the Rhinog Formation (defined on the basis of the presence of abundant sandstone beds) is probably slightly diachronous between the two sections.

There are considerable lateral variations in bed characteristics between the two sections- this is partly due to the complex amalgamated nature of the beds and as a result a lack of clearly defined packets. The lateral persistence of the Thin Bedded Facies (91-100m at Foel Wen, 43-53m at Craig y Fechan) is striking and suggests that this facies may be laterally persistent over several kilometres and thus has a tabular geometry, in contrast to the more laterally variable Sand-rich Facies.

ii) <u>Base of the Barmouth Formation.</u> (Figs 4.64, 4.65)

Logs were measured from Garn [SH 616 165], Gell Fawr [SH 621 175], Bwlch y Llan [SH 620 178], Bwlch Cwm Maria [SH 621 190], Llyn Irddyn [SH 635 226], and Llyn Dulyn [SH 660 241]. The base of the Barmouth Formation is sharp and probably synchronous between all the sections. The only section where the stratigraphic position is uncertain is Llyn Irddyn, where the top of the Hafotty Formation was not exposed. The base of the exposed sandstones at Llyn Irddyn may be correlated tentatively with the base of the Barmouth Formation, with 28m in the Bwlch Cwm Maria section or perhaps 30m in the Garn section (equivalent to 38m in the Bwlch Cwm Maria section. The former is the most likely correlation. Fig 4.65 highlights the probable lateral persistence of the Thin Bedded Facies, though there appear to be lateral differences in the depths to which they have been eroded and then filled by overlying deposits.

As in previous examples lateral changes occur in bed characteristics and although general trends are similar between logs there are many important differences. Some of the greatest differences occur between Garn and Bwlch Cwm Maria, the former containing Thin Bedded Facies between 50-55m which is relatively laterally persistent (occurring

FIG 4.64 Base of the Barmouth Formation logs: distribution.

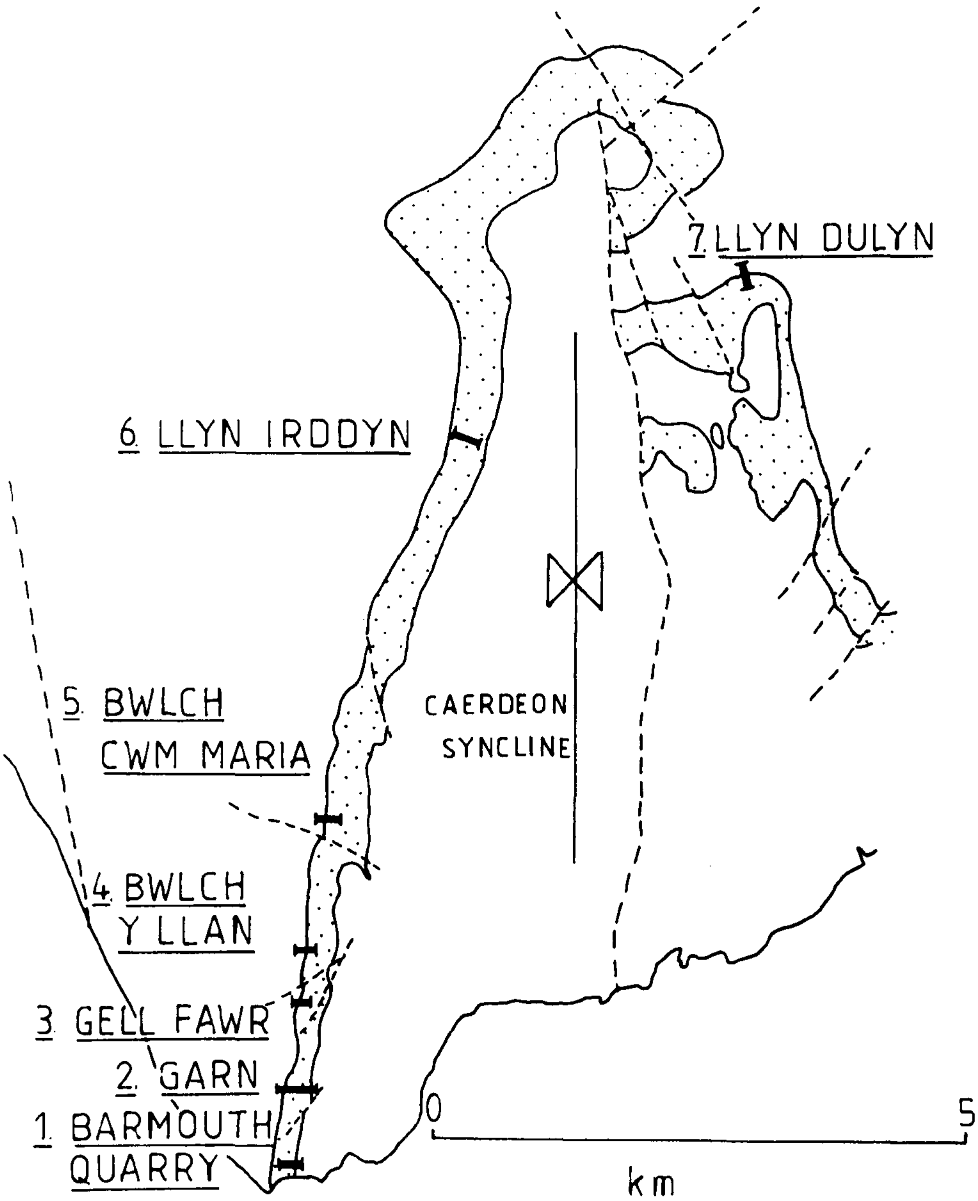
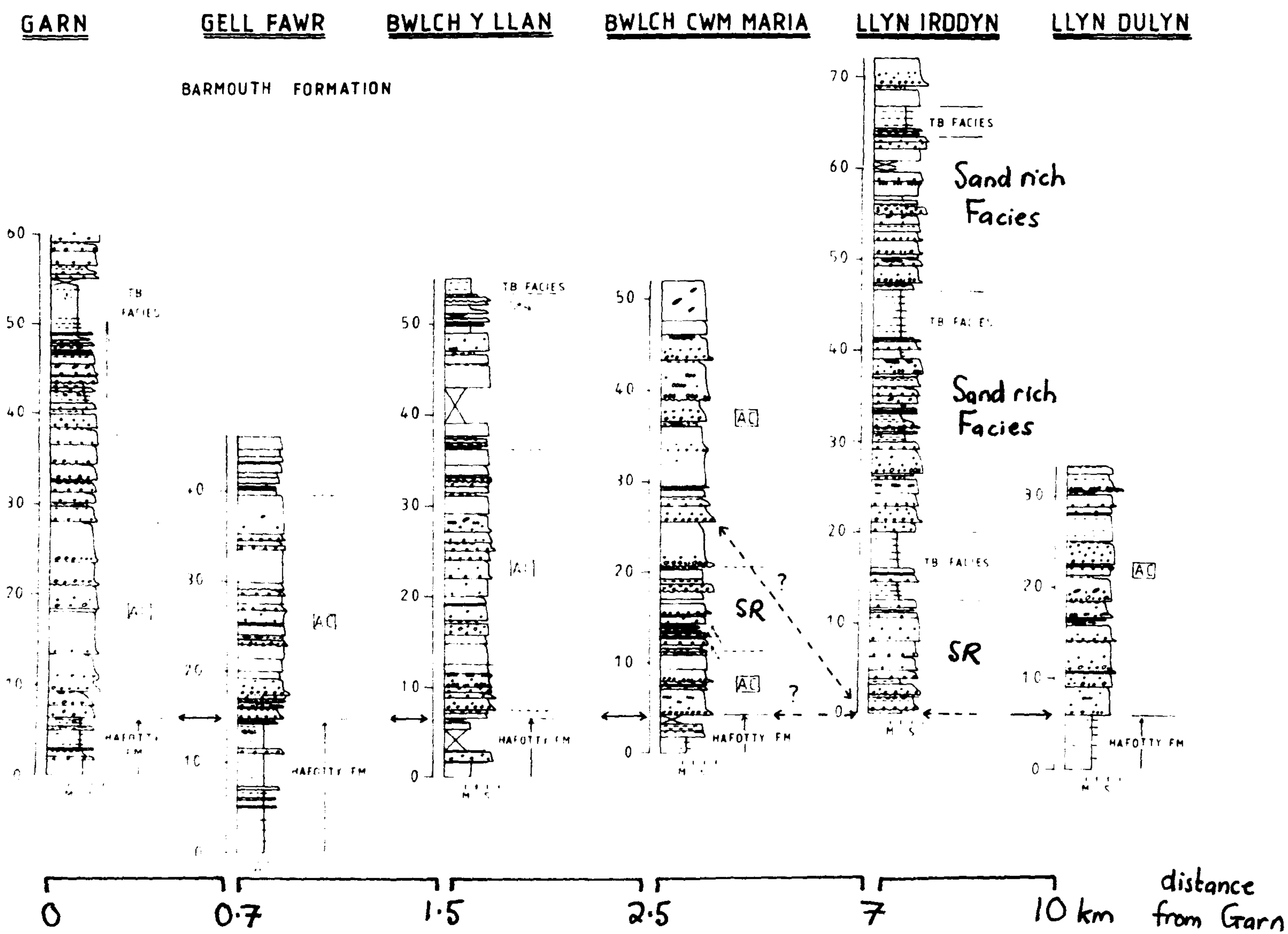


FIG 4.65 Base of the Barmouth Formation Logs.



at Bwlch y Llan and possibly Llyn Irddyn), while at Bwlch Cwm Maria at the same stratigraphic horizon there are thick bedded coarse sandstones. The fact that these sandstones are rich in intraclasts implies that they may have been erosive into Thin-Bedded Facies. The logs are quite different in other parts of the sequence, for instance thinner bedded units at Bwlch Cwm Maria pass laterally into Amalgamated Coarse grained Facies at Garn.

The common occurrence of Amalgamated Coarse Grained Facies and the abundance of amalgamated beds in general are therefore associated with great lateral variability in bed characteristics. The lack of clear definition of sand packets mean that assessing larger scale geometries is difficult. There do not appear to be any consistent lateral changes, though there may be at least some channelisation.

iii) <u>Top of the Barmouth Formation.</u> (Fig 4.66)

Despite the appearance of a tabular geometry to beds as seen for example in Barmouth Quarry, there is considerable variation in bed characteristics between Barmouth Quarry [SH 616 155] and Garn [SH 616 165] (about 1 km distance), though sandstone packets appear to be relatively tabular. Interesting differences occur; the thick, amalgamated sandstone bed at 48m at Barmouth thins markedly towards Garn (155m), while some 50cm thick graded beds (c.35m, Barmouth) amalgamate northwards to form a thick massive amalgamated bed above 140m at Garn. There are few consistent lateral changes apart from a slight coarsening towards Garn.

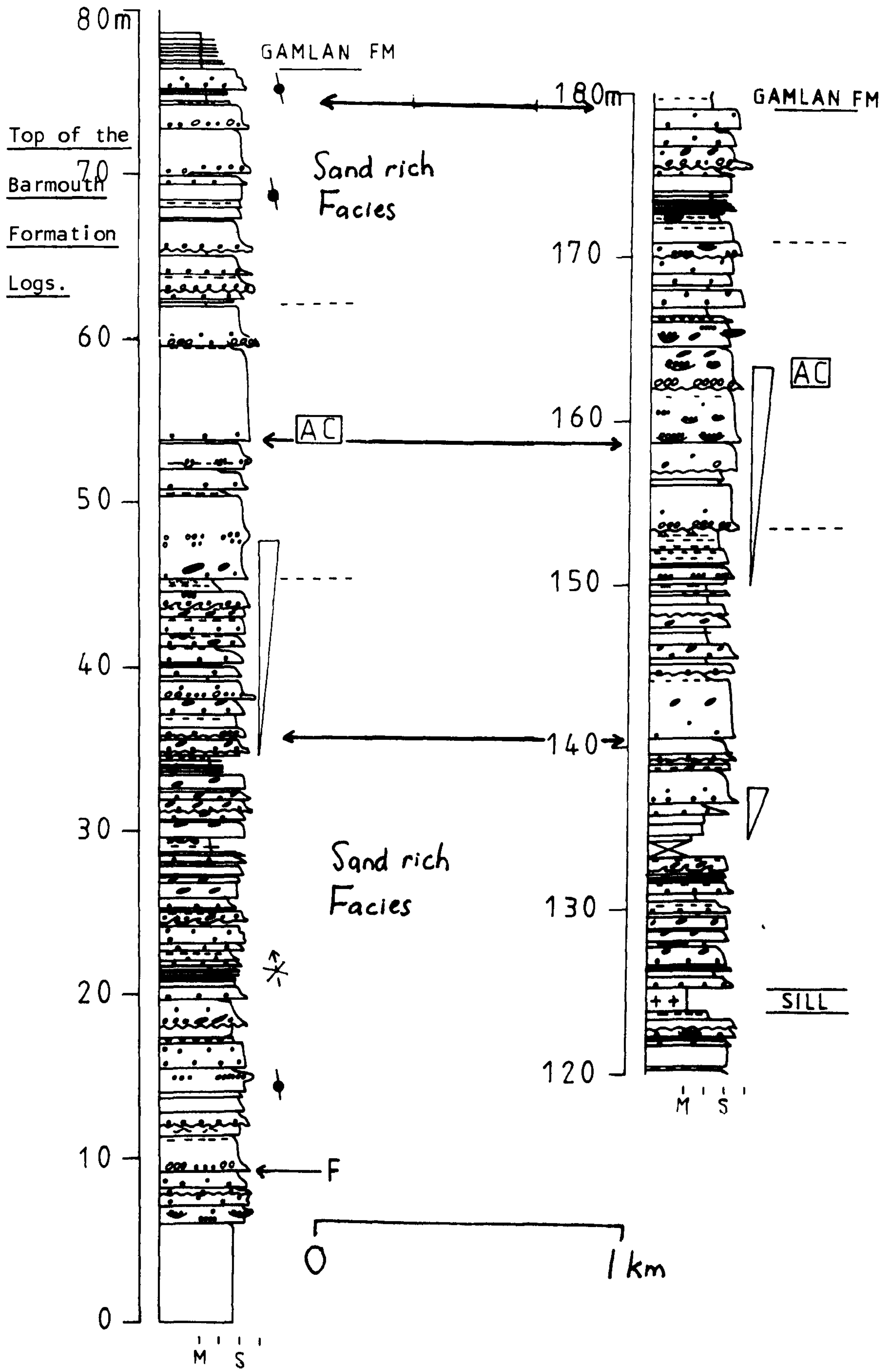
4) <u>Harlech Dome Scale.</u> Lateral variability over tens of kilometres.

Few beds or packets can be traced laterally on this scale, though the Manganese Grit, the Bluestone Grit and the Cefn Coch Grit have a sufficiently tabular geometry for them to be correlated across much of the Dome (see section 4.10, the Hafotty Formation). Although some formations show lateral thickness changes, possibly due to the diachronous

BARMOUTH QUARRY GARN

FIG 4.66

BARMOUTH FORMATION



nature of some of the facies, overall most have a sheet-like geometry.

Assessment of Bed/packet Geometry.

In general on the outcrop scale, apart from the Amalgamated Coarse Grained Facies and areas associated with abundant cross-bedding, beds are usually tabular (e.g Clogwyn Pot, Barmouth Quarry). However beyond this scale beds are lensoid and are very difficult to follow for long distances (i.e greater than 1km). This is particularly the case where sandstone beds are amalgamated, where individual turbidites, the products of single events, may be difficult to separate.

Packets of sandstone and Thin-Bedded Facies appear to be more laterally continuous and thus more tabular in geometry (for instance if one views the southwest face of Craig Wion, east of Cwm Bychan which is approximately 750m long). Some sediment packets can be correlated on the kilometre scale e.g the Manganese Grit and these clearly have a sheet-like geometry, though within the Rhinog and Barmouth Formations assessment of the packet geometry is more difficult. Sand packets can be regarded as broad (kilometre scale) sheets. How sand packets lense out laterally is not clear- it appears to be relatively gradational, though whether thinning takes place on upper bed or packet surfaces or lower surfaces, i.e as an upstanding lobe or an incised channel, is not clear. There seems to be evidence for both on the bed scale e.g F Rock, section 1, though in this case the compensating effects of the sedimentary system led to the production of relatively tabular packets of sediment. The geometry of units within the Rhinog and Barmouth Formations is therefore complex.

4.8 : Vertical Sequences.

The analysis of ancient turbidite sequences relies heavily on the interpretation of vertical thickness and grain size changes, since often individual turbidite beds are not well enough exposed for their large-scale geometry to be determined. Thus vertical trends up through the stratigraphic sequence have been used to determine sub-environments on submarine fans, to decide whether progradation or aggradation are important in particular turbidite systems and to identify major changes in supply to the turbidite system (for a review see Walker 1978; Ricci Lucchi 1984). Different controls on sedimentation produce vertical changes at different scales.

It is therefore important to evaluate whether there are any trends within the Rhinog and Barmouth Formations or any evidence for cyclicity. The Rhinog Fawr section will be used as an example section through the Rhinog Formation; it is the thickest sequence that was looked at in detail and comprises a substantial proportion of the overall formation thickness. The Garn section will be used as an example of a sequence in the Barmouth Formation since it is a relatively continuous section through virtually the whole formation. As well as these logs, the thickness of each sandstone bed was plotted against its position stratigraphically within the sequence. The mean bed thickness and the sandstone-siltstone ratio were calculated at 10m intervals up through the sequence. Where both Thin Bedded Facies and Sand-rich Facies were present in a given grouping the mean value and the individual values for each facies are given (open circles). The Roman Steps section (Rhinog Formation) was also examined in order to look at the smallest scale, bed-by-bed trends.

1) Rhinog Fawr Section. (Figs 4.67, 4.68, 4.69)

Large Scale. In general, as one goes higher up in the sequence the Thin Bedded Facies decreases in volumetric

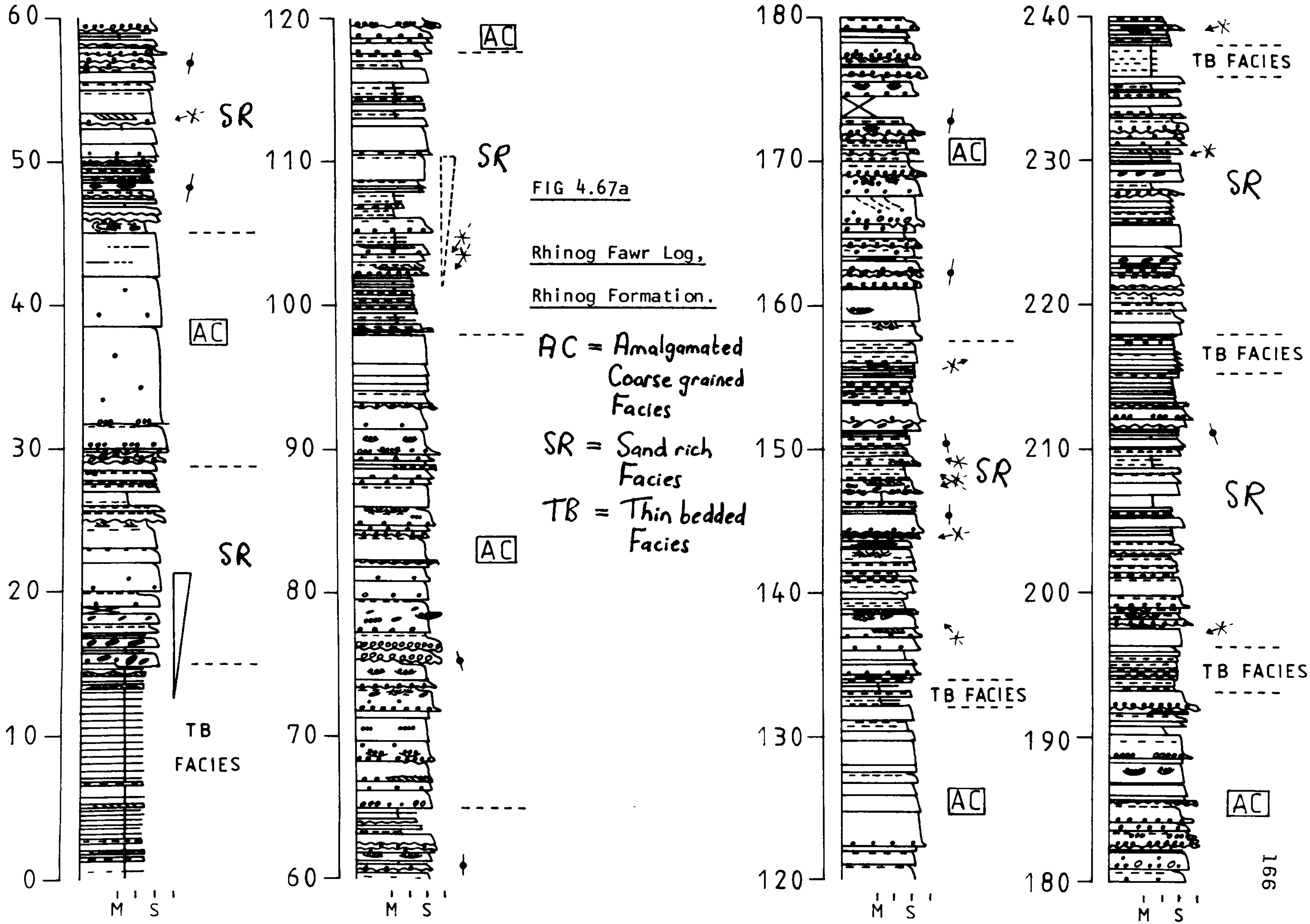


FIG 4.67a

Rhinog Fawr Log,

Rhinog Formation.

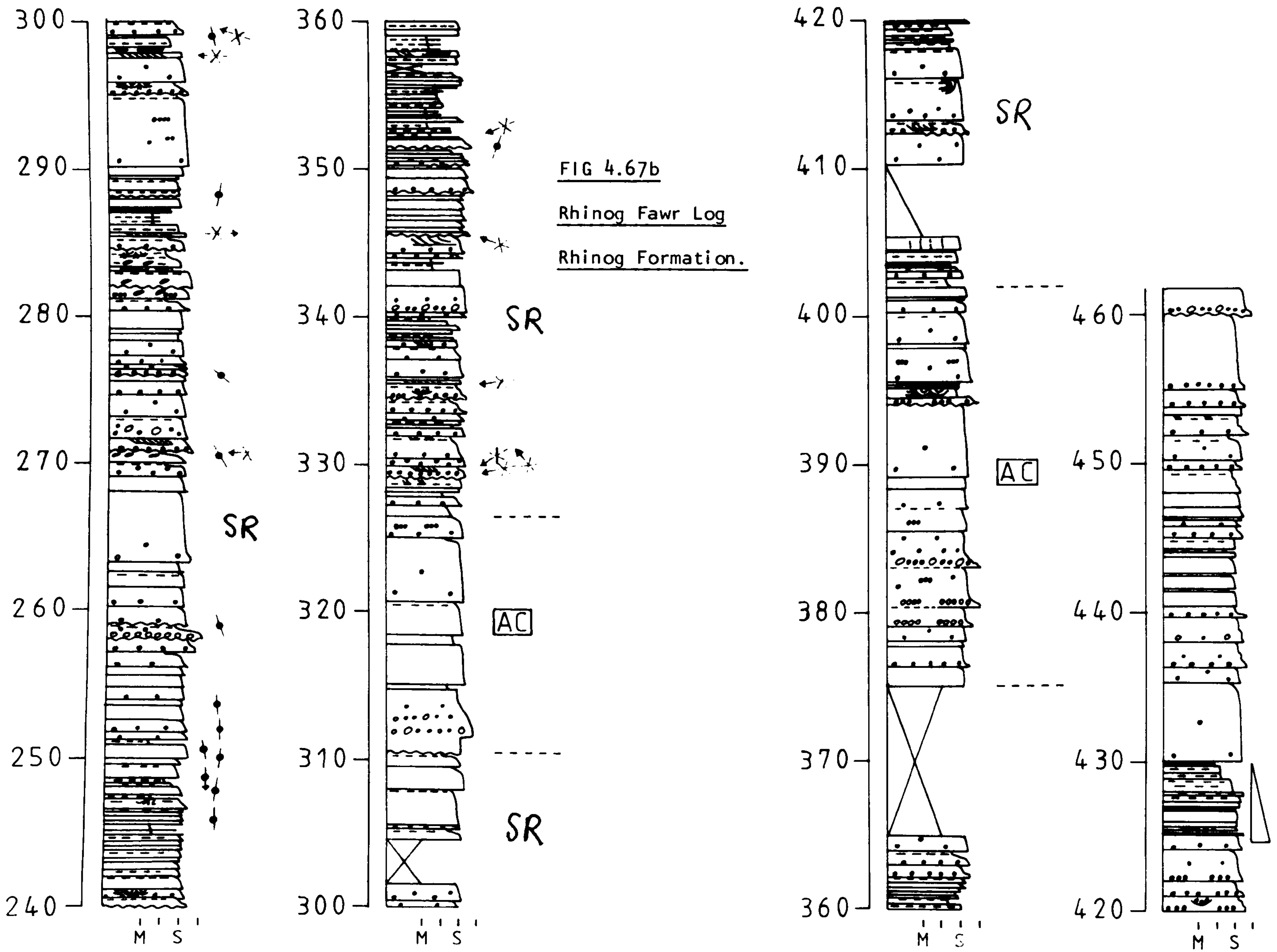


FIG 4.67b

Rhinog Fawr Log

Rhinog Formation.

FIG 4.68 A plot of Bed Number (counted from the base) against Bed Thickness, Rhinog Fawr section, Rhinog Formation.

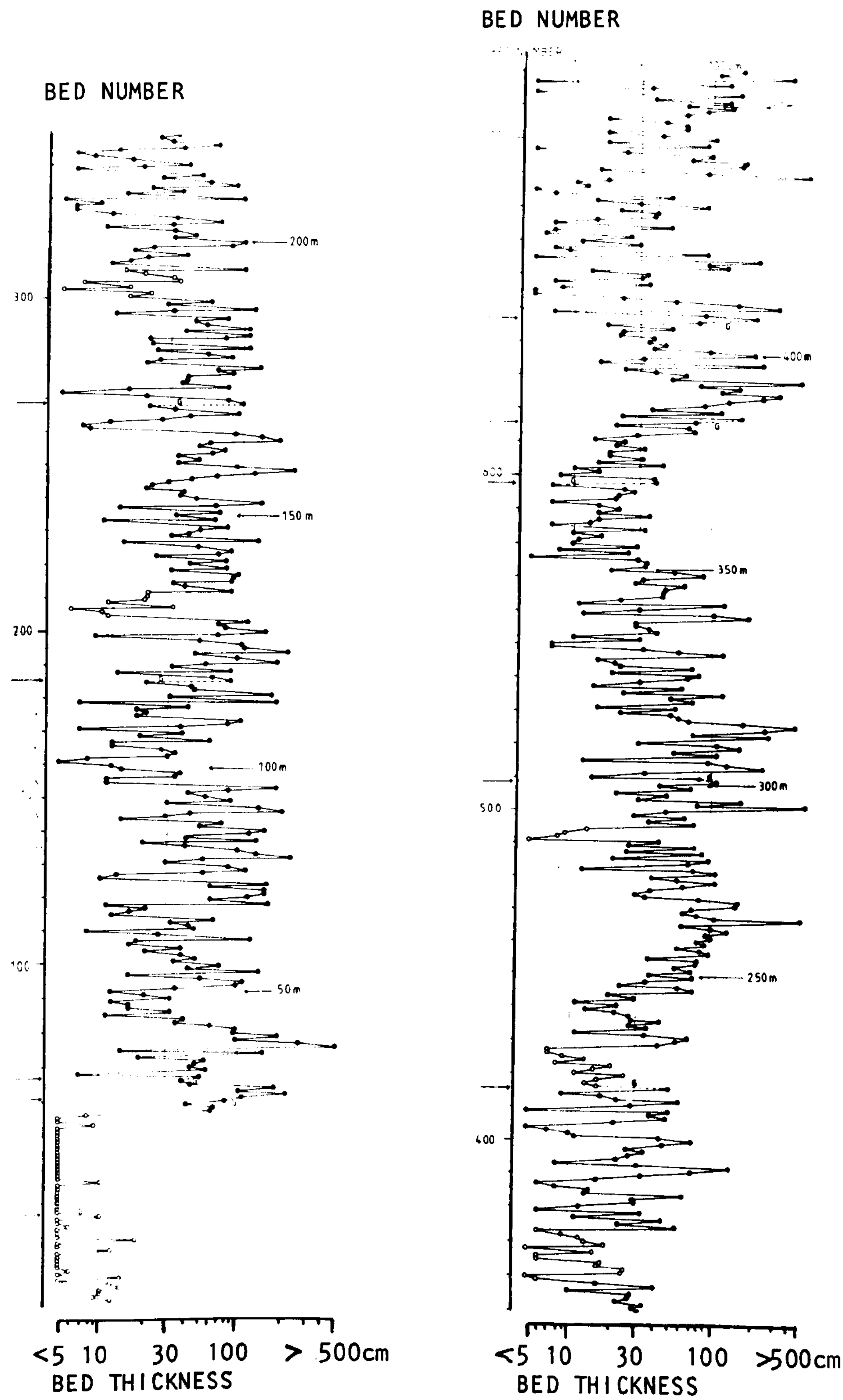
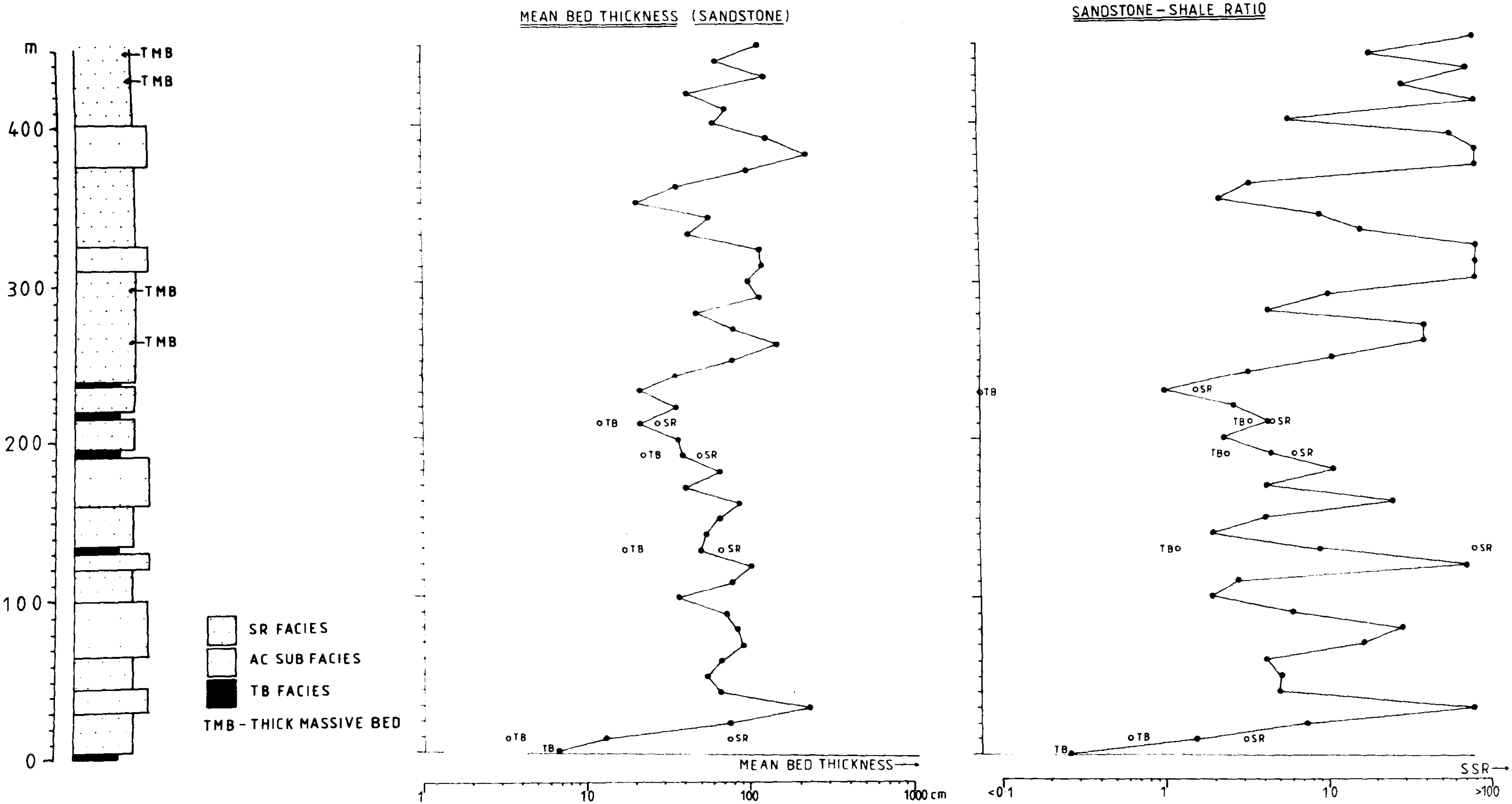


FIG 4.69 A plot of Facies, Mean Bed Thickness and Sandstone-Shale Ratios, Rhinog Fawr section, Rhinog Formation.



importance, the mean bed thickness and the sandstone-siltstone ratio increases slightly. These trends seem to be due to a reduction in the frequency of thinner beds higher up the sequence rather than a tendency for beds to get thicker relative to beds lower in the sequence. If only the Sand-rich Facies is considered then the overall change in mean thickness is very small, though the increase in sandstone-siltstone ratio is slightly more pronounced. This large-scale change may be due to progradation of the turbidite system, though this was probably of a limited nature since in general there are no major facies changes or evidence of upward coarsening through the sequence.

The sequence does however seem to fall into two major parts, with a division at 230m, which is particularly apparent on the sandstone-siltstone ratio diagram (Fig 4.69). The lower part of the sequence shows after an initial (12-20m) increase in bed thickness and sandstone-siltstone ratio (?progradation of the turbidite system) an overall decrease in these characteristics (30-230m), though with medium-scale ?cycles superimposed. The level of correlation is insufficient to enable the lateral extent of these trends to be determined, but the large-scale nature of these trends suggests that they may be source-controlled (c.f. Lajoie 1979). The upper part of the sequence (above 230m) is more irregular and determination of overall trends is much more uncertain.

<u>Medium Scale.</u> On this scale judgement of trends is highly subjective. However in Fig 4.68 (bed number versus bed thickness) the general trend in bed thickness appears to increase and decrease in the shape of a sine wave about the 30cm bed thickness line, with a wavelength of approximately 100 beds. This cyclicity may possibly be related to roughly symmetrical thickening and thinning sequences, each of the order of 50 beds. Most discrepancies to this cyclicity occur where there are large gaps in the exposed sequence or may be related to amalgamated sequences, where determination of units produced by individual events may be difficult. Mean bed thickness show similar trends with alternating higher

and lower bed thicknesses about the 70cm mean thickness line, with a wavelength of approximately 40-50m. There are also alternations of higher and lower sandstone-siltstone ratios on a similar scale, though the cycles may be variable (having a range of 20-60m). Superimposed on this cyclicity there may be larger scale trends (see above).

The cycles described above can be correlated with facies and/or subfacies changes. In general the parts of the sequence with the highest bed thicknesses, mean bed thicknesses and sandstone-siltstone ratios correlate with the Amalgamated Coarse Grained Facies, thick massive beds and the thicker bedded, coarser grained parts of the Sand-rich Facies. Thinner bedded sequences with low mean bed thicknesses and sandstone-siltstone ratios correlate with the Thin Bedded Facies and the finer grained parts of the Sand-rich Facies. Approximately 11 cycles may be detected and 3 types can be differentiated:

i) Asymmetric with gradual increases in bed thickness and relatively abrupt tops, e.g. 0-20m is well defined, 220-260m is transitional with the symmetrical type (iii). These appear to be progradational, being commonly underlain by Thin Bedded Facies.

ii) Asymmetric with relatively abrupt bases to the sequence and gradational tops. This type may be the most important type within the overall fining upward sequence 30-230m, on the 20-30m scale.

iii) Symmetrical cycles. On the 40-50m scale this appears to be the most common type, if the slightly asymmetric cycle is also included, e.g 280-350m.

As well as these cycles determined from bed thickness and sandstone-siltstone ratios additional fining and thinning and coarsening and thickening trends can be defined from the log (Fig 4.67). These trends are subjectively defined; they occur on a variety of scales and rarely show consistent direction, e.g. stacked coarsening and thickening upward sequences.

Small Scale- (e.g Fig 4.67). On the small scale (less than 10m), vertical trends appear to be more irregular. In some parts of the sequence there are 5-10 bed alternations and there may also be 2 bed alternations superimposed.

2) Garn Section- (Figs 4.70, 4.71)

Large Scale- The Barmouth Formation has sharp upper and lower boundaries and although the thickest beds occur near the base and the coarsest beds are present in the lower part of the succession, there are no clear formation wide trends.

Medium Scale- There appear to be 3 or 4 cycles of variable thickness (40-60m scale) which fit into the 3 types discussed above:

i) 120-180m is a well developed asymmetric sequence.

ii) 0-50m is a good example of an asymmetric sequence and may be interpreted as a channel fill (see later) and contains Thin Bedded Facies at the top of the sequence.

iii) 60-120m is a symmetrical sequence.

Small Scale- Both fining and thinning upward (e.g. 40-50m) and coarsening and thickening upward sequences (e.g. 65-75, 150-160m) occasionally occur on the scale of approximately 10m. There is a slight tendency for Amalgamated Coarse Grained Facies to be associated with coarsening upwards sequences and Thin Bedded Facies to occur in fining upward sequences. However the junctions of these facies are usually relatively sharp.

3) Roman Steps Section. (Fig 4.72)

The Roman Steps Section was examined in some detail in order to attempt a more precise appraisal of bed-by-bed vertical trends. This sequence occurs in approximately the middle part of the Rhinog Formation. The sequence can be divided on the basis of facies and subfacies.

Two facies occur within the sequence: Thin Bedded Facies (0-2 and 46-7m) and the Sand-rich Facies. There is also a transitional part of the sequence between 47-50m,

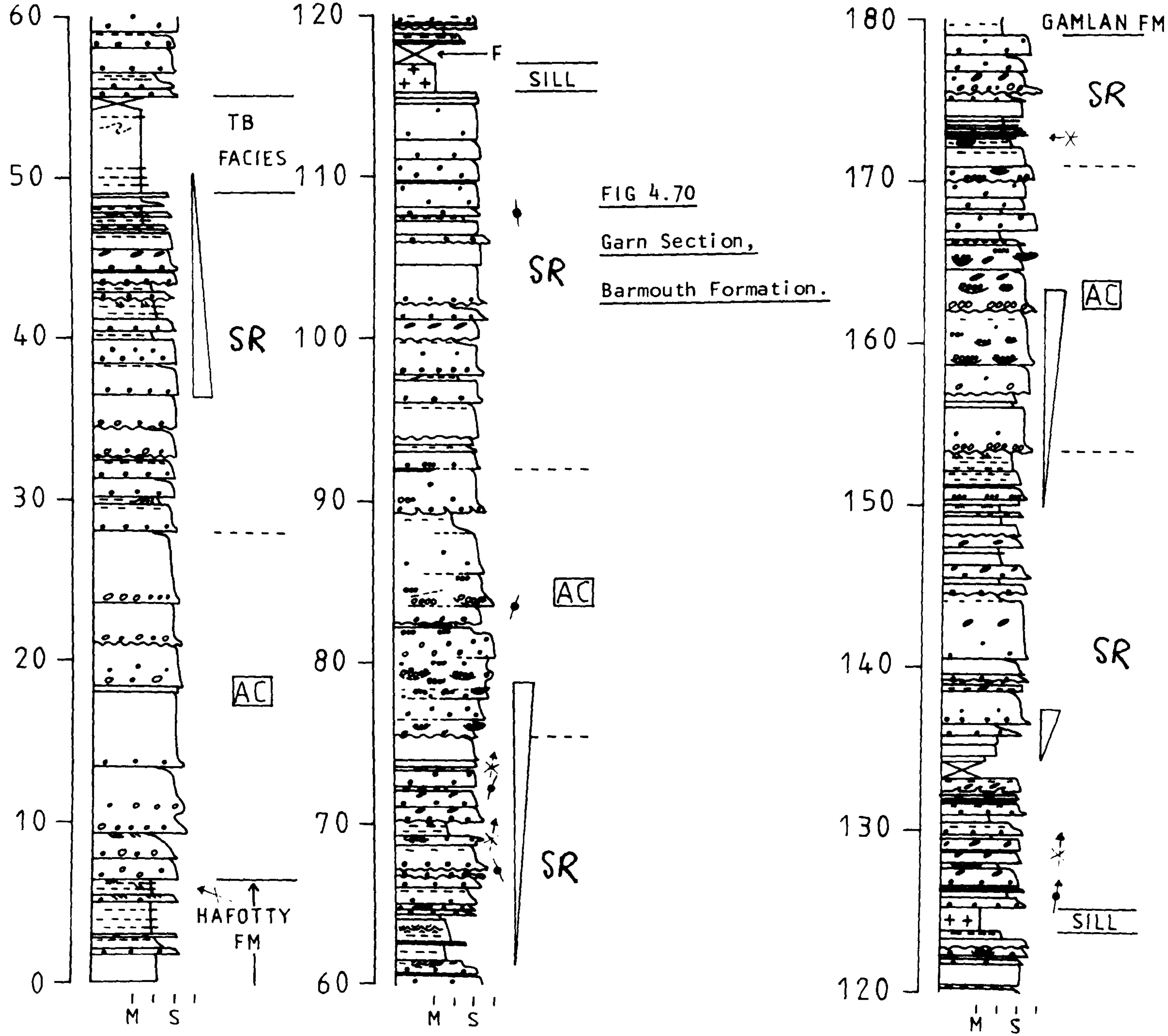


FIG 4.70

Garn Section,

Barmouth Formation.

FIG 4.71 A plot of Facies, Mean Bed Thickness and Sandstone-Shale Ratios, Garn section, Barmouth Formation.

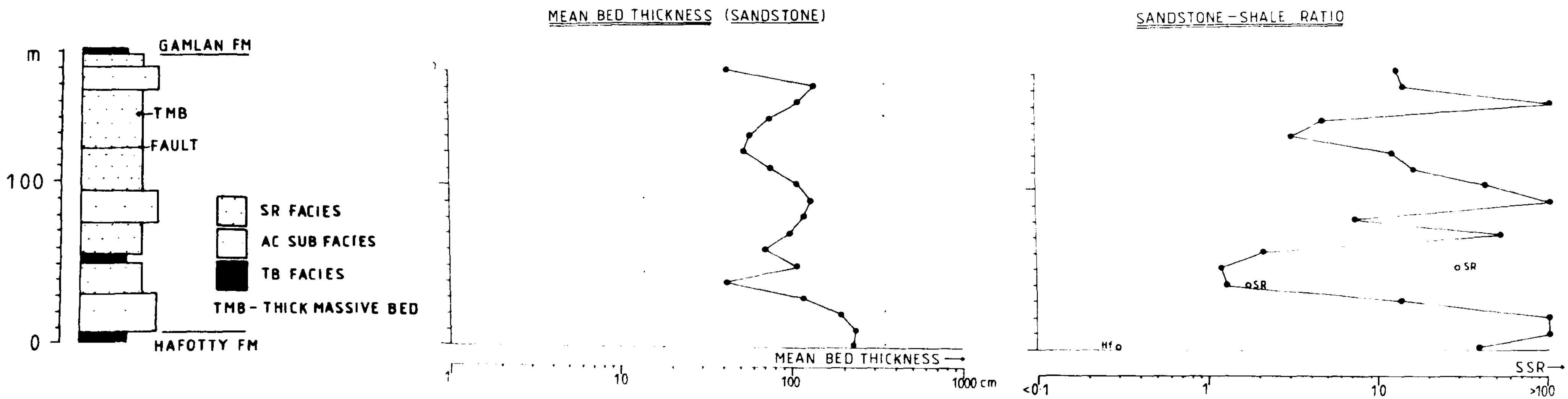
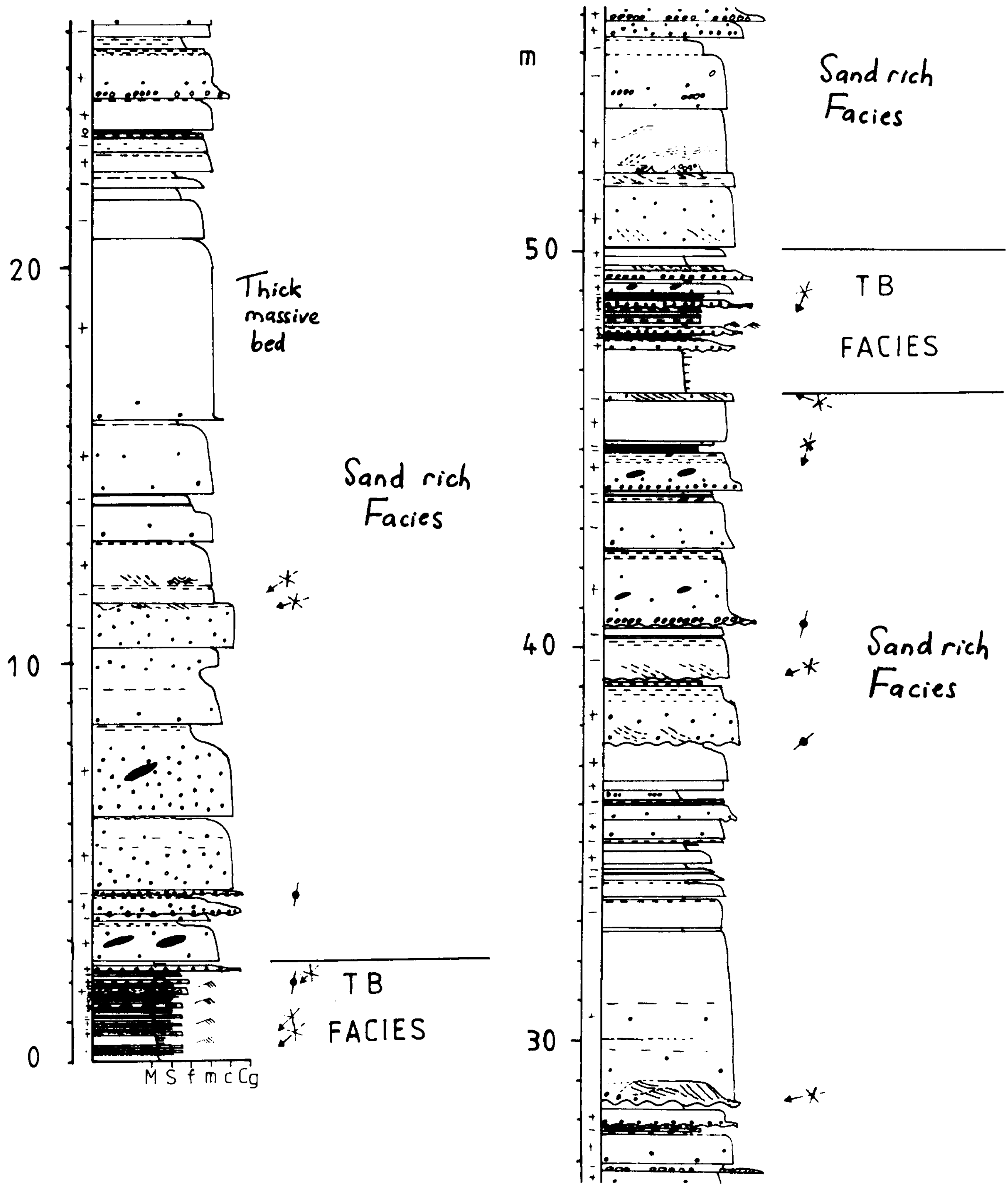


FIG 4.72 Roman Steps Section, Rhinog Formation.



which has a similar average bed thickness to the Thin Bedded Facies (13cm) with a range of 1-24cm. However as well as containing some base absent turbidites (e.g. T_{be},T_{ce}) this sequence also contains abundant relatively coarse grained graded beds, often overlain by parallel lamination and/or ripple cross lamination (e.g. T_{abe}, T_{ae}, T_{ace}). In this case it is difficult to rigidly define facies and correspondingly makes interpretation of vertical sequences more difficult.

As shown in Fig 4.72 it is possible to subdivide the sequence of Sand-rich facies on the basis of bed thickness variations and other bed characteristics using an adaption of the sub-facies scheme described earlier in the chapter:

i) Thin bedded graded subfacies (occurs at 22-28m, 33-36m and is transitional with the Thin Bedded Facies 47-50m). Mean bed thickness commonly varies from 27-42cm, range 5-126cm, i.e. it includes some thick beds. Sandstone-siltstone ratio ranges from 2 to 5, which is higher than the Thin Bedded Facies but is in general lower than other parts of the Sand-rich Facies. It is largely composed of thin bedded, graded T_{abe}, T_{ae} with some T_{be} (type C1 beds, Mutti 1979) and is only occasionally amalgamated.

ii) Medium/Thick bedded subfacies (occurs 2-16m and 36-46m). Mean bed thickness commonly varies from 50-100cm, range 1-230cm; sandstone-siltstone ratio is about 20. Most beds show good grading (mainly normal grading) as well as cross-bedding as isolated sets. T_a, T_{ab}, and T_{abc} sequences are the most common, some of which contain S1,2 and 3 divisions. Amalgamation is common (type C1 and B1 of Mutti).

iii) Medium/Thick bedded subfacies, similar bed thickness characteristics and high sandstone-siltstone ratios but composed of massive beds showing only poor, discontinuous grading. If grading is present it is restricted to the bases of beds, where flame strucures and possible dewatering features are also present. This type is indicative of Mutti's type B1 beds and may indicate flows of even higher density than the type "ii" beds.

iv) Thick Massive Beds (16-21m and 28-33m). Both are about 400cm thick, are poorly graded and massive, similar to bed type "iii", though they may be associated with cross-bedding e.g. at 28m. They are almost certainly amalgamated but the similarity in grain size makes it impossible to distinguish separate flow events.

It is important to look at the sub-facies boundaries. Bed types typical of certain sub-facies sometimes occur interbedded with dissimilar bed types. Thus bed types are not necessarily mutually exclusive within a given facies or sub-facies. However some facies/sub-facies boundaries are sharp, especially the upper boundary of the Thin Bedded Facies. Small scale sequences may occasionally occur e.g. the fining upward sequence 33-35m, but there are no consistent changes up through the sequence and subfacies are relatively well defined with relatively abrupt boundaries. Other sequences may however show more transitional boundaries (e.g. Rhinog Fawr Section).

Interpretation.

Trends of differing type occur at different scales in vertical sequences within the Rhinog and Barmouth Formations. Although these trends in bed thickness and sandstone-siltstone ratio can be related closely to the facies the lack of consistent clear directional sequences makes interpretation difficult. These sequences therefore do not display stacked fining and thinning upwards sequences (typical of turbidite channel fill) or stacked coarsening and thickening upward sequences (typical of repeated lobe progradation). Consistent trends appear to be absent and so in this case vertical sequences cannot simply be used as direct indicators of environments on a turbidite fan. The relationship of these sequences to environmental interpretation is discussed more fully in section 4.13.

4.9 : Trace Fossils.

A range of trace fossils are found in the Rhinog Formation. They occur at the following localities (Fig 4.73, solid squares represent *in situ* occurrences of trace fossils, open circles- loose blocks):

1) [SH 6632 3057] northeast of Llyn Morwynion, south of the 518m spot-height.

2) [SH 6655 3106] southeast side of F Rock.

3) [SH 6670 3132] north of F Rock.

4) [SH 6617 3143] Valley bottom, southwest side of stream, northwest of F Rock.

5) [SH 6649 3103] south of F Rock, loose block found in the scree.

6) [SH 6571 3012] north of the Roman Steps.

7) [SH 6625 3118] west of F Rock.

8) [SH 6610 3160] Valley bottom, northeast side of stream, northwest of F Rock.

9) [SH 6614 2958] Lower northeast slopes of Rhinog Fawr, loose block.

Most trace fossils in the Rhinog Formation occur in the Thin Bedded Facies or more rarely in the Sand-rich Facies. This may suggest that the trace fossils were produced during periods when the sedimentation rate was lower and their preservation potential was higher. The occurrence of trace fossils indicates that the bottom waters were sufficiently oxygenated to permit life.

Hypichnial trace fossils i.e. those which occur at the base of sandstone beds are the most common, and occur as sandstone-filled casts of burrows and possibly some trails. In the field they are found most commonly where the underlying finer grained bed has been weathered away to leave convex downwards, positive relief casts. These bedding planes are often stained light brown (?limonitic). In general trace fossils are relatively uncommon; they are

FIG 4.73 Map of trace fossil localities, Rhinog Formation.

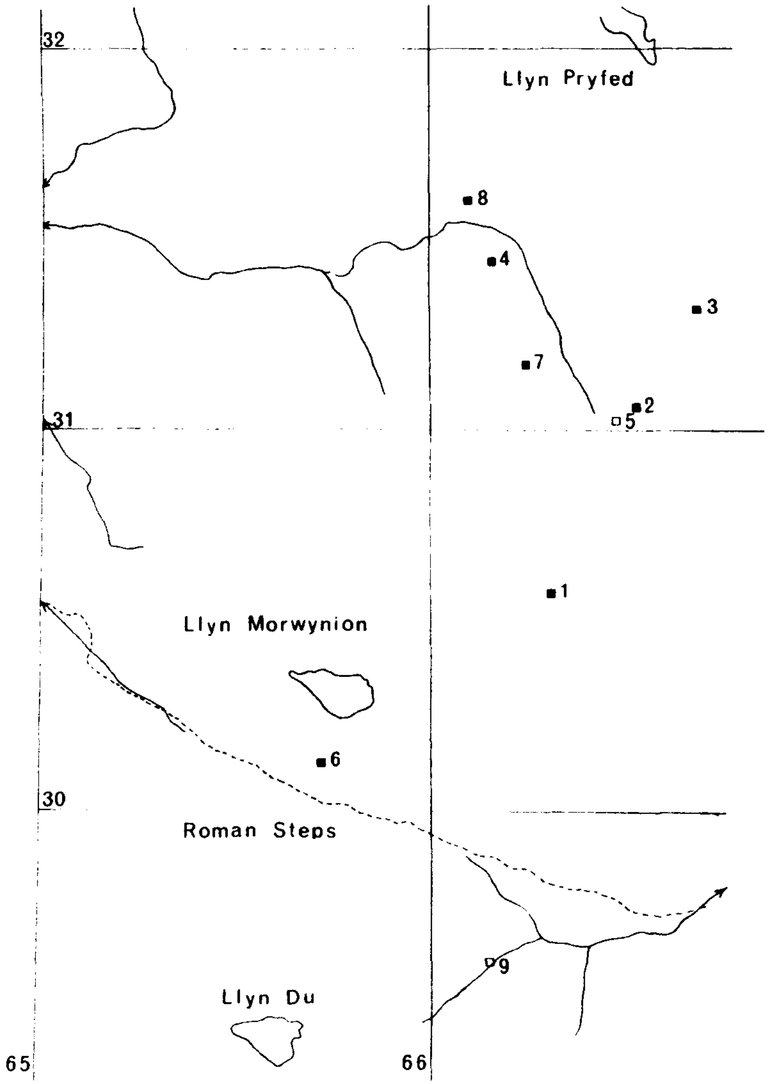
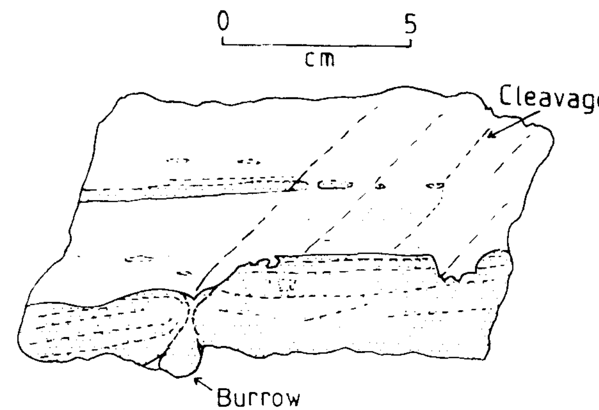


FIG 4.74 Cross section of a large burrow, Rhinog Formation.



restricted to certain bedding planes even within the Thin Bedded Facies and only at locality 4 are they generally abundant.

Infaunal burrows predominate. These can be divided into two main types- horizontal or bedding parallel burrows and vertical or bedding discordant burrows.

Horizontal Burrows.

Although these burrows vary greatly in size most are between 0.5 and 5mm wide. Their length is very variable; usually there is a positive correlation between burrow width and length. Burrows about 0.5mm wide can usually be followed for up to 1cm, whereas burrows which are about 4mm wide are usually up to 10cm long. Measured lengths may only be regarded as minimum burrow lengths since often parts of individual burrows run oblique to the bedding plane. The horizontal burrows can be divided into the following size ranges:

i) Small. 0.5-1mm wide, up to 4cm long.

ii) Medium. 2-3mm wide, often irregular in shape, can be 5cm or more long.

iii) Large. 4-5mm wide, often very long, up to 20cm.

iv) Very large. Greater than 5mm wide. The largest burrows were found at locality 4; these are approximately 2cm wide and 25cm long (Plate 4/XV). They are smooth walled and bifurcate.

Another very large burrow is shown in Plate 4/XVI and its cross section in Fig 4.74. This burrow is 12-15mm wide, 8mm deep and 15cm long. It has been infilled by fine sandstone associated with the bed above. The underlying mudstone has been weathered away apart from two thin bands next to the burrow. The sandstone bed is 2cm thick, but it is clear that above the trace fossil the bed thins and has been disrupted. This may have resulted from either preferential dewatering at the site of the trace fossil, or deformation processes (either soft sediment or tectonic). The disruption resulted in thinning of the sandstone bed above the trace fossil and cleavage is now concentrated in

PLATES 4/XV to 4/XX : Trace Fossils from the Rhinog Formation.

PLATE 4/XV

PLATE 4/XVI

PLATE 4/XVII

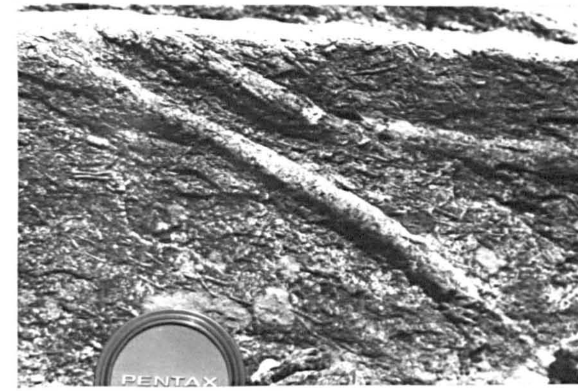


PLATE 4/XVIII

PLATE 4/XIX

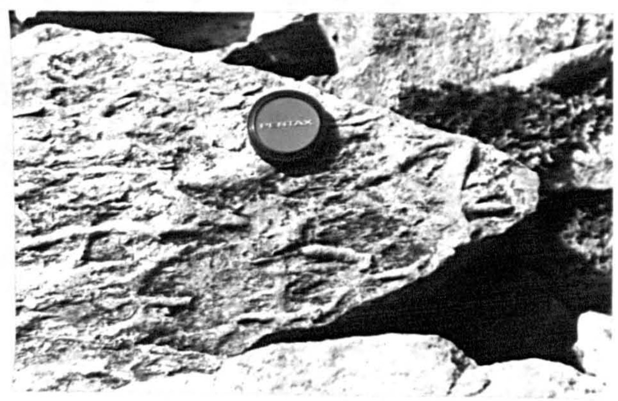
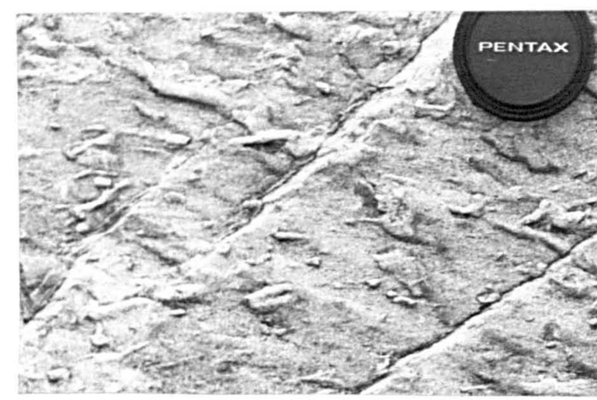


PLATE 4/XX



this area. In the upper part of the bed laminae have been truncated by epichnial burrows and a thin graded lamina above has been disrupted by bioturbation.

Trace fossils of different size often occur together. In the above example type "iv" horizontal burrows occur with type "ii", as well as some vertical burrows. Thus there may be a strongly bimodal size distribution of burrows on a given bedding plane. What controls the size of burrows is unclear, the only evidence being that generally smaller burrows tend to be rarer in coarser sediment, where larger burrows predominate. Larger organisms (producing larger burrows) probably found it easier to burrow in coarser sediment. Burrows usually remain a constant width, though rarely some show variations.

The burrows are usually entirely filled with sediment, but in some types there is a central hole to the fill of the burrow, usually less than 0.5mm wide. One type was observed in longitudinal section which had two cavities on either side of a central sediment filled axis. Most burrows have smooth walls, though rare examples show irregular surfaces.

The burrows may be straight (which is more common in the larger burrows), slightly sinuous (the most abundant type) or highly sinuous (which is more common in the smaller burrows). It is noticeable that some burrows run parallel to each other, and rarely they form horizontal U-shaped burrows. Plate 4/XVII shows a probable small U burrow on the left hand side of the photo (3cm above the lense cap). There is also a strong bimodality in the size of the trace fossils, with large straight burrows and small burrows- some of which bifurcate. Some burrows also appear to be bilobate, which may be an original feature of a single burrow, a tight U-burrow or a compacted single burrow.

A relatively common feature of some of the burrows is the presence of bifurcations. Burrows may bifurcate parallel to the major burrow, or the burrow may bifurcate at an acute angle in a dendroid pattern (Plate 4/XVIII). The bifurcations usually occur in a consistent direction, but they may occur rarely in different orientations. Locally burrowing may become so intense that burrows may link up to form network structures with polygonal inter-burrow areas (Plate 4/XIX at locality 7). Burrows may also radiate from a common centre to form dense patches of horizontal burrows, for instance a small radiating group of horizontal burrows is seen to the right of the lense cap in Plate 4/XVIII.

Some of these trace fossils show a greater degree of organisation than would be expected from simple deposit feeding organisms (which possibly constructed the simple horizontal burrows- *Planolites*). Instead the bifurcating "clumped" burrows are typical of *Phycodes* and possibly *Chondrites* where dichotomous bifurcations are more common. *Phycodes* was a dwelling burrow, produced by organisms with a suspension feeding mode of life. The horizontal U-burrows may be likened to *Rhizocorallium*, though no spreite were found associated with them. The network structures may even suggest a foraging or farming mode of life for their producer, by analogy with the similar network system of *Palaeodictyon*.

Vertical Burrows.

Vertical burrows usually range from 0.5-4mm in width, with a mean of 1mm. They are very common on some bedding planes and are often associated with horizontal burrows. Vertical burrows and horizontal burrows often have a similar width and locally horizontal burrows turn, producing vertical or oblique burrows. There are four main types of vertical burrow:

i) Large vertical burrows between 4-20mm across, are often oval in cross section. This may represent an oblique section through a large burrow or the original shape of the burrow.

ii) Vertical burrows with a circular or near circular cross section. This is differentiated from type iv by being unpaired (*Skolithos ?linearis*).

iii) Vertical burrow, which is annulated. One example was found; the burrow is 2mm wide and has horizontal striations on its outer surface at 0.5mm intervals

(*Skolithos annulata*).

iv) Paired vertical burrows, suggesting that they were connected as part of a vertical U burrow. The distance from one entrance of the burrow to another varies from 1-10mm, most commonly being 3-4mm. Since no spreite were found this may be assigned to the ichnogenus *Arenicolites*.

The relative age of the burrowing is problematical in many cases. In Plate 4/XX predominantly vertical paired burrows occur on the base of the bedding plane, along with flutes and other scour features. Some of the burrows are definitely post-depositional (i.e. produced after the deposition of the sandstone bed which infilled the scour structures) since they cut some of the flute casts produced immediately prior to deposition. However in other cases flutes seem to nucleate on scoured vertical burrows suggesting that the burrows may have formed bed topography favourable for flute formation (i.e. producing flow separation in the vicinity of the burrow).

Many of the trace fossils described above could be assigned to the shallow water *Skolithos* or *Cruziana* facies of Seilacher (1967) and Crimes (1975). However Crimes (1977) and Crimes *et al.* (1981) have described similar trace fossils which occur in deep sea fan environments. Crimes (1977) suggested that similarities between some shallow water and inner to mid fan environments produced similar infaunal behavioural responses. Thus there is no conflict between the occurrence of apparent shallow water trace fossils and an interpretation of the Rhinog Formation as being deposited in a turbidite basin.

The relative lack of diversity of trace fossils is partly a reflection of the environment of deposition- the sand-dominated substrates would be nutrient-poor and thus unfavourable to benthic and burrowing faunas. Trace fossils most commonly occur where the Thin Bedded Facies forms relatively thick sequences, representing periods of reduced turbidite erosion and deposition, when the top few centimetres of the sediment column directly below the sediment-water interface was comparatively nutrient-rich. Crimes (1974) has also suggested that although there was rapid diversification of shallow water ichnogenera in the Lower Cambrian, few forms have been found in deeper water deposits. Deeper water environments were not extensively colonised until after the Cambrian and this may also account for the lack of "deep water" trace fossils in the Lower and Middle Cambrian of North Wales.

4.10 : Hafotty Formation.

The Hafotty Formation has an average thickness of 150-200m and a maximum thickness of 250m (Matley & Wilson 1946, Rushton 1972). The formation can be divided into three units (Fig 4.75):

i) The Lower or Ore-bed shales (15-20m thick), which contain a manganese ore-bed, overlain by manganiferous blue-grey mudstones (the "Bluestone") and a sandstone bed (the "Bluestone Grit") (Woodland 1939). A log through the Ore-bed shales near Harlech [SH 5921 3189] is shown in Fig 4.76.

ii) The Manganese Grit: a sandstone-rich packet (0 to 60m thick), which can be recognised over much of the Harlech Dome.

iii) Upper Shales (100-200m thick).

Most of the Hafotty Formation is made up of thinly bedded siltstones and mudstones, usually green or grey in colour and well cleaved. Base absent Bouma sequences are common and sandstone-siltstone ratios are generally low. Occasional parallel laminated sandstone beds are present (T_b) as well as ripple cross lamination within T_{bc} and T_{cd} sequences, typically 10-20cm thick. The cross laminated divisions do not normally exceed 4cm thick and the cross laminae in general indicate flow towards the west. Thicker, coarser beds occur rarely; some are well graded and contain intraclasts. These beds are up to 50cm thick and occur as T_{ae} beds containing grains of coarse sand and finer. Coarser beds are slightly more common towards the top of the Hafotty Formation (for instance at Garn [SH 617 167] though sandstone-siltstone ratios are low even in this part of the sequence.

In the northeastern Harlech Dome the Hafotty Formation is more sandstone-rich than elsewhere. The sequence at Craig-y-Penmaen [SH 727308], for example, contains sandstones up to 32cm thick which are graded at the base, have parallel laminated tops (T_{ab}) and include some

FIG 4.75 Schematic log through the Hafotty Formation.

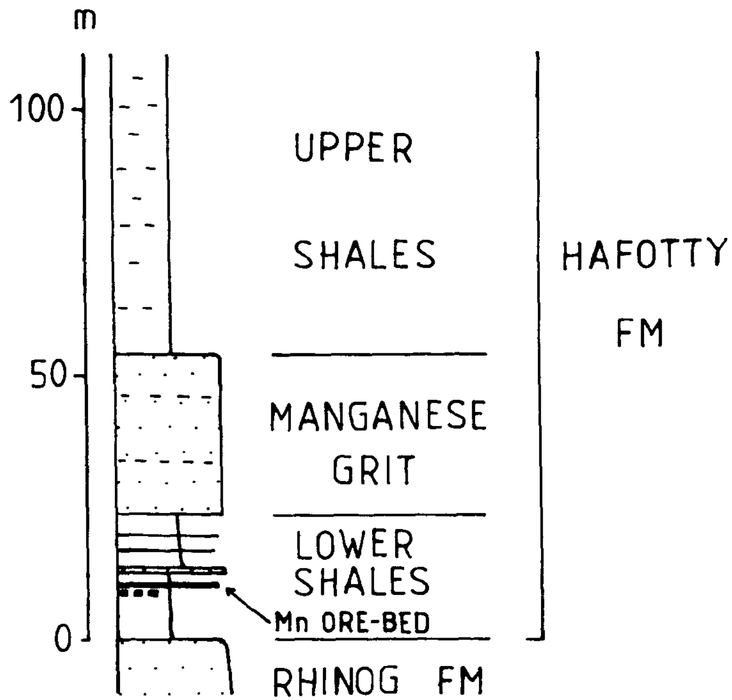
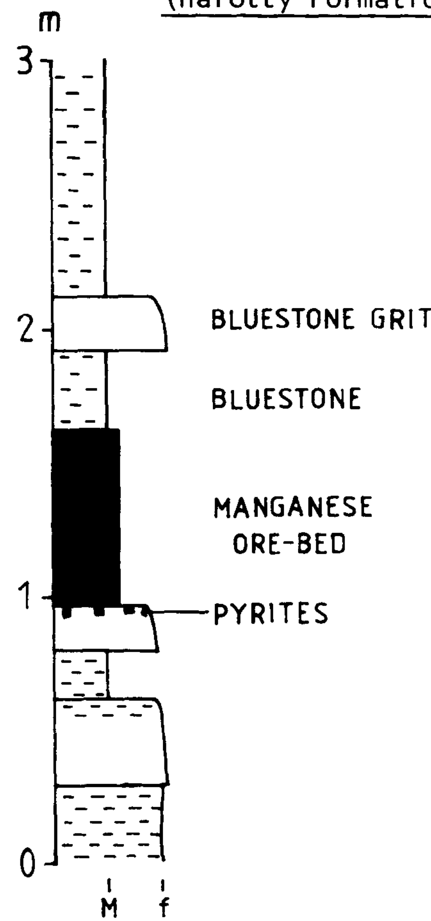


FIG 4.76 Log through the Ore-bed Shales (Hafotty Formation), at Harlech.



T_{abc} units. However sandstone beds are on average 3-10cm thick, and are parallel laminated (T_{be}) and/or cross laminated (T_{bc}, T_{cd}) with cross-laminae indicating flow towards the west-north-west. Occasional convolute laminated beds also occur.

In general this facies resembles the Thin Bedded Facies of the Rhinog Formation and some bed types in the Gamlan Formation. The Hafotty Formation was also deposited from dilute turbidity currents. No sole structures which would indicate the palaeocurrent direction were found, though the presence of a sandier facies in the northeastern Harlech Dome might suggest a source in that direction. However the presence of coarse grained, thicker bedded T_a units near the top of the Hafotty Formation, directly below the Barmouth Formation which is derived from the south, may indicate that some sandstone beds in the Hafotty Formation were also derived from the south.

The facies described above is the most abundant in the Hafotty Formation. However the characteristics and geometry of three particular units are important in determining the depositional setting of this formation:

i) Manganese Grit- an example of a laterally continuous sandstone packet, which is comprised of Sand-rich Facies.

ii) Bluestone Grit- a laterally continuous bed of sandstone.

iii) Manganese Ore-bed.

<u>Manganese Grit</u>

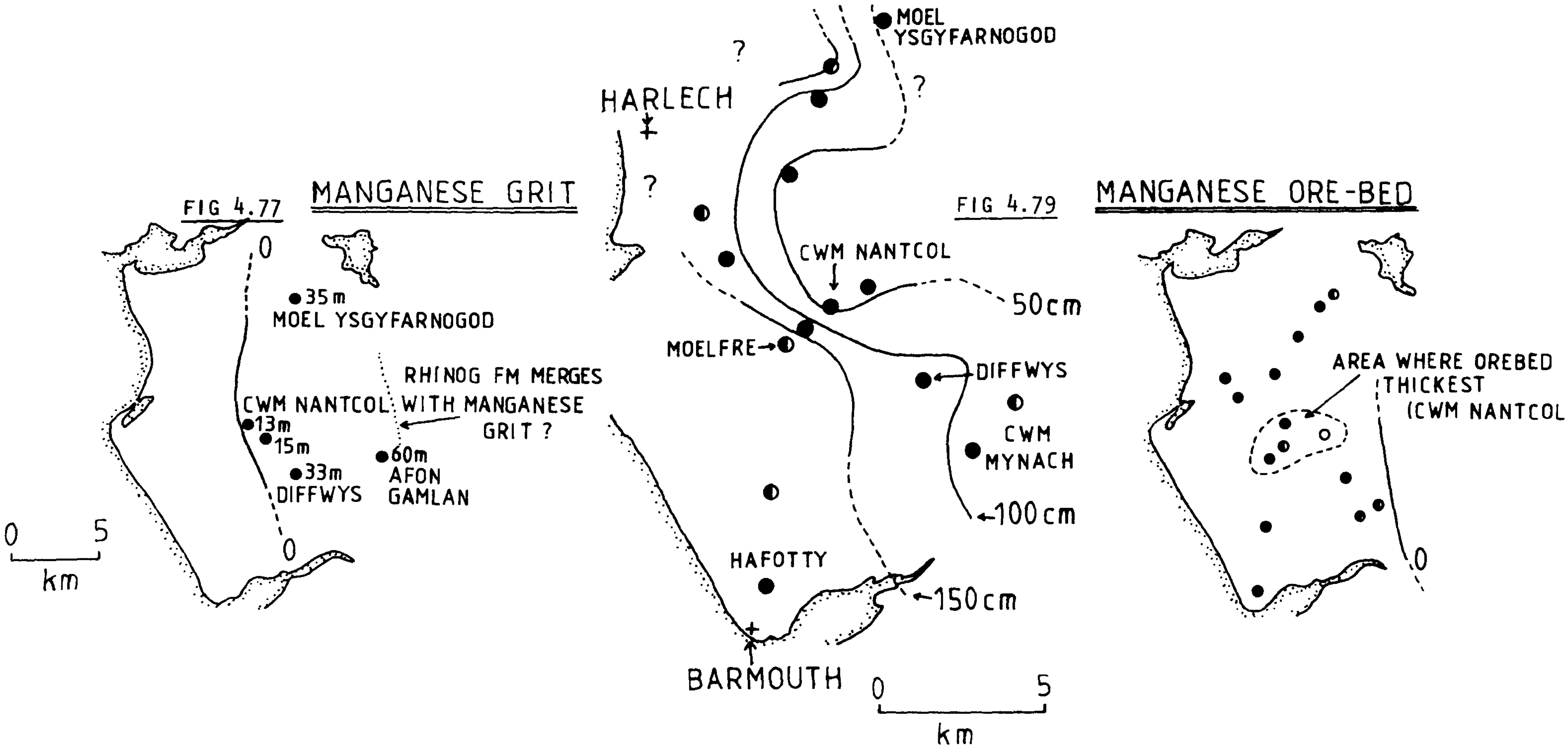
The Manganese Grit (so-named because of its proximity to the Manganese Ore-bed not to its manganese content) is a unit of amalgamated sandstones, which ranges from 0-60m thick, being thickest in the northern Harlech Dome and thinnest towards the south-west (Fig 4.77). The Manganese Grit has not been identified on the eastern side of the Harlech Dome, north of Afon Gamlan, where the Manganese

Isopachytes for the:

FIG 4.78 BLUESTONE GRIT

FIG 4.77 MANGANESE GRIT

FIG 4.79 MANGANESE ORE-BED



Ore-bed is also missing. Here the sandstone beds of the Manganese Grit may lie immediately above the Rhinog Formation and so would not be recognisable as a separate unit. Regionally the Manganese Grit thins rapidly westwards in a zone running approximately NNW-SSE between Foel Wen, Moelfre and the Caerdeon Syncline fold axis (Fig 4.77). To the northeast of this line the Grit forms a thick mappable horizon, thickening towards the east; it is 13m thick at Cwm Nantcol, 33m at Diffwys and 60m in Afon Gamlan (present observations plus data from Matley & Wilson 1946).

The Manganese Grit is typically thick to very thickly bedded. It may be poorly bedded with few sedimentary structures as at Moel Ysgyfarnogod [SH 6583 3415], or thickly bedded with 1-2m thick amalgamated sandstones as at Foel Penolau [SH 6622 3473]. Scour structures with pebble fills are locally common and some are very large e.g. near Clip [SH 657 333]. Parallel lamination may occur rarely near the top of medium sandstone beds. Other sedimentary structures are uncommon apart from occasional grading and flame structures e.g. at Craig Uchaf [SH 6491 2687]. The Manganese Grit is therefore comprised of amalgamated Bouma T_a and rare T_{ab}, top absent sequences, which indicate deposition by high density sand-rich turbidity currents.

No unequivocal palaeocurrent indicators were found, though some rare scours indicate a N-S flow orientation. Isopachytes (Fig 4.77) might suggest that the Manganese Grit was derived from the east since it thickens in that direction. The stratigraphic proximity of the Manganese Grit to the Rhinog Formation, the latter of which is derived from the north, their lithological similarities and the presence of N-S oriented scours may suggest a similar source, i.e. to the north.

Thus after a lull in coarse clastic deposition during which the Manganese Ore-bed was formed, the Manganese Grit was deposited as closely spaced pulses of sediment, possibly determined by one type of process such as a sequence of major slump failures developing into turbidity currents, or as a related sequence of events such a cluster of seismic shocks. The turbidites form a single package of sediment that is continuous over much of the Harlech Dome, in an area at least 16km by 10km. This degree of lateral continuity may be analogous to Mutti *et al.*'s (1984) "megaturbidites" which are basin-wide and seismically induced. The Manganese Grit does not have a basin-wide distribution, but it appears to have been largely unconstrained while being deposited and may have a similar origin.

If the Manganese Grit was deposited from turbidity currents resulting from slope failure, essentially at a single site, then individual beds might be laterally variable in thickness, but overall the packet could have a tabular geometry, lensing out only at the flow margins (Fig 4.77- the zero isopachyte). If on the other hand, the Grit was formed from turbidity currents originating from several sites along the basin margin, possibly triggered by seismic shocks, then an amalgamated, sheet-like deposit might also be produced. The different point sources would probably not be detectable in the absence of palaeocurrent indicators. It is not possible to detect the subtly different geometries of individual beds and thus is difficult to evaluate which hypothesis is correct. However the rapid thinning of the Manganese Grit to the west may tentatively favour the single source hypothesis. The N-S alignment of isopachytes probably indicates flow from the north, the zero isopachyte representing the westward limit of the turbidite lobe.

Bluestone Grit

The Bluestone Grit occurs above the Manganese Ore-bed and in most places is a single bed of sandstone grading up into sandy mudstone. This bed is laterally continuous throughout most of the Harlech Dome. Thickness variations were described by Woodland (1939) and from this data it is possible to construct approximate isopachytes (Fig 4.78). The Bluestone Grit is divided into two sandstones by the presence of a thin bed of mudstone (in Fig 4.78 these are denoted by the half-filled circles), but the distribution of this mudstone parting is irregular. The Bluestone Grit is coarsest in the east and south, though it is thickest in the west, thinning and fining towards the centre of the Dome where it is only 30-40cm thick. To the north-east near Moel Ysgyfarnogod the Grit is coarser than in the central part of the Dome, though it is also thinner (Woodland 1939). Thus the Bluestone Grit has a complex thickness and grain size pattern which is not easily explained. However there seems to be an inverse relationship between the thickness of the ore-bed and the thickness of the Bluestone Grit.

Manganese Ore-Bed

The most distinctive feature of the Hafotty Formation is the presence of a manganese ore-bed. The ore-bed occurs over most of the Harlech Dome, except in the north-east part of the Dome, north of Afon Gamlan; its absence here may be due to greater clastic supply from this direction. The ore-bed can be correlated with manganiferous shales (Mulfran Beds) on St Tudwal's peninsula (Nicholas 1915).

On average the ore-bed is 30cm thick and is composed of a very fine grained intergrowth of spessartite garnets and rhodochrosite (manganese carbonate) as well as some disseminated quartz (Woodland 1939, 1956). Overall the ore-bed has a manganese content of 37-49%. The ore-bed is hard, cherty, resistant to cleavage and is composed of alternating red (with hematite inclusions), pink, yellow and bluish black bands (with pyrolusite).

The ore-bed is concordant with beds above and below and in most places is underlain by a pyritic mudstone and merges above with the Bluestone (a sequence of finely laminated manganiferous blue-grey mudstones). This evidence combined with the constancy in thickness, mineralogy and geochemistry of the ore-bed over the Harlech Dome support a sedimentary origin for this stratiform ore deposit (Woodland 1939).

The absence of detrital or organic grains and the euhedral nature of some the spessartite garnets argue against a primary detrital origin for the ore-bed. The uniform grain size (fine silt), irregular banding, presence of spheroids- some of them compacted in the plane of the banding, presence of Liesegang ring structures, patchy nature of the cream coloured ore (colloform type?) and the occurrence of synaeresis cracks led Woodland (1939) to suggest that the ore-bed was deposited chemically as a colloidal gel, composed of rhodochrosite and clay with some siliceous material. The original sedimentary/diagenetic textures are still preserved but during chlorite grade regional metamorphism some of the rhodochrosite converted to spessartite (Woodland 1939).

Two main hypotheses have been proposed for the origin of the ore-bed:

i) Depositional Origin.

The average Mn/Fe ratio of continental crust is 1:50 (Roy 1976). However the Manganese Bed has a ratio of 7.9:1 (Mohr 1964). It is therefore necessary to explain not only the abundance of manganese but also the concom itant lack of iron. The Hafotty Formation was fed either by a source rich only in manganese or there was selective leaching of manganese at source, manganese being generally more mobile. Thus it would be likely that mainly manganese would be deposited because it was predominantly manganese that was supplied to the basin. However this hypothesis would require extremely unusual chemical conditions to allow such large amounts of manganese to be transported (e.g low pH, anoxic) and deposited directly.

Another possibility is that manganese was concentrated within the basin of deposition rather than at source. This scenario would require an enclosed or silled basin (Demaison & Moore 1980), to allow manganese to concentrate. Although the average manganese content of seawater is low (0.002 ppm), in the presence of hydrogen sulphide large amounts of manganese can be dissolved, as in the case of the enclosed, partially anoxic, stratified Black Sea, where 100 million tons of fluvially-supplied manganese are at present in solution (Roy 1976). The Hafotty Formation shows little evidence for widespread anoxic conditions within the water column (such as black organic-rich shales), though the pyrites-rich mudstone may indicate temporary anoxic conditions. In the Black Sea model manganese is concentrated relative to iron since the first minerals to precipitate are usually the sulphides (in seawater with a relatively low pH). Iron sulphide is precipitated first (the pyrite-rich mudstone possibly represents this event), but manganese sulphide does not form since it is too unstable. Manganese would precipitate later in seawater with a higher pH, possibly due to overturning or upwelling of the water column (Roy 1976).

Bennett (1987) suggests that manganese may have been supplied to the basin by submarine exhalative hydrothermal solutions.

ii) Diagenetic Origin.

During periods of high sedimentation and reducing conditions there may be extensive remobilisation of manganese within the sediment column and fractionation of manganese relative to iron.

In this case dissolved manganese concentration takes place in the upper few metres of the sediment profile. Under conditions of high bicarbonate activity rhodochrosite may precipitate close to the sediment-water interface. The accumulation of manganese-rich sediments in Loch Fyne (Scotland) and the Baltic Sea (Glasby 1974) offer potential modern analogues for the accumulation of the Hafotty ore-bed and suggests that under ideal conditions sufficiently high dissolved manganese concentrations can be attained, allowing pore waters to reach chemical equilibrium with respect to manganese carbonate (e.g. Lynn & Bonatti 1965; Calvert & Price 1972).

The source of bicarbonate necessary for rhodochrosite precipitation has been discussed by Coleman *et al.* (1982). Under conditions of shallow burial diagenesis in marine environments two potential sources of bicarbonate are available (Irwin *et al.* 1977):

i) dissolved marine bicarbonate and

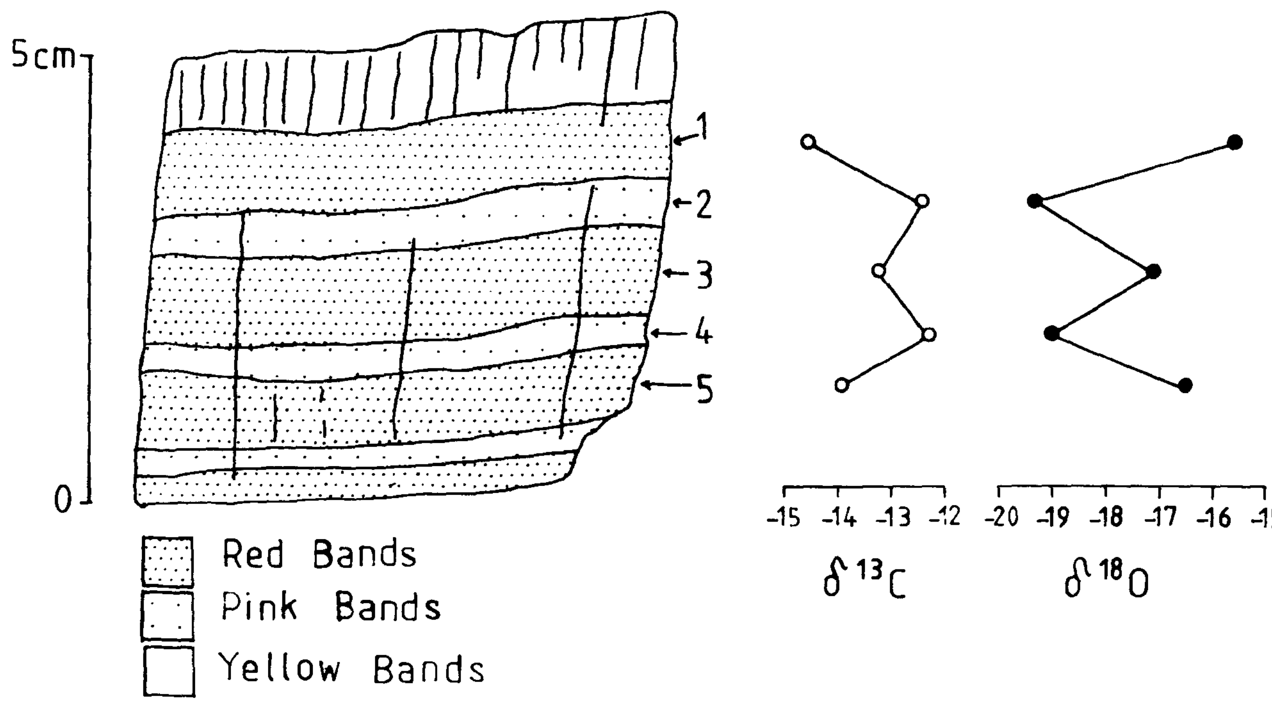
ii) bicarbonate derived from the bacterial degradation of organic matter.

Potentially the source of bicarbonate can be resolved on the basis of their distinctly different stable isotopic compositions (Irwin *et al.* 1977). Generally speaking dissolved marine bicarbonate has a carbon isotopic composition of 0 per mil while bicarbonate produced by the bacterially induced decomposition of organic matter induces extreme isotopic fractionations (Irwin *et al.* 1977).

In order to shed light on the possibility of a diagenetic origin for the ore-bed an exploratory carbon and oxygen stable isotopic study of a single banded sample from Moel Ysgyfarnogod [SH 658 337] was undertaken using analytical facilities at the University of Liverpool (see appendix 1 for analytical details). X-ray diffraction showed rhodochrosite to be the only carbonate phase present within the sample. Whole rock analysis (based on five replicate samples) gave a mean carbon and oxygen isotopic composition of -12.93 per mil and del -17.43 per mil respectively. Sub-samples taken separately from alternating red bands and pink (light) bands show an inverse relationship between carbon and oxygen values (Fig 4.80). The red bands have more depleted del ^{13}C values (mean= -14.01) than the pink bands (mean=-12.45). Conversely the pink bands have more depleted del ^{18}O values (mean= -19.10) than the red bands (mean=-16.29). X ray diffraction analyses show that the alternating bands are mineralogically similar, although the red bands contain slightly more spessartine and slightly less rhodochrosite than the pink bands.

The highly depleted isotopic composition of the ore-bed rhodochrosite reported here is significantly different from that expected for precipitation from Cambrian seawater with an approximate value of 0 per mil and -5 per mil for del ^{13}C and del^{18}O respectively (James & Choquette 1983). Given that the Lower Palaeozoic sequence as a whole has been subject to regional low-grade prehnite-pumpellyite to lower green-schist metamorphism (Bevins & Rowbotham 1983) it is difficult to accept that the isotopic values reported

FIG 4.80 Variations in Carbon and Oxygen isotopes, Manganese Ore-bed, Hafotty Formation, Moel Ysgarnfogod.



here represent primary compositions. Oxygen isotopic data clearly reflect the effects of high temperature metamorphism. First approximation temperatures calculated on the basis of the temperature controlled fractionation of oxygen isotopes between the solid and fluid phase give temperatures of approximately 300 ± 50°C. These calculations assume a value for metamorphic fluids of approximately +6 to +8 per mil (SMOW), a reasonable value for low grade metamorphic fluids (Taylor 1979) and use the temperature controlled fractionation equation for calcite-water (Friedman & O'Neill 1977) which should approximate reasonably to the rhodochrosite-water fractionation equation (which has yet to be defined).

Compared to oxygen isotopes, carbon isotopic compositions are much less prone to change by recrystallisation processes due to low concentrations of dissolved inorganic carbon held in solution (Buchbinder *et al.* 1984). The del ^{13}C values reported here for rhodochrosite fall within the range of values which might be expected for carbon generated by bacterially mediated sulphate reduction occurring in anoxic sediments (Irwin *et al.* 1977). The occurrence of a pyrite-rich mudstone directly beneath the ore-bed indirectly supports the development of sulphate reducing conditions. If sulphate reduction was the dominant process during the formation of rhodochrosite then the development of sulphate reducing conditions would be ideal for the concentration of manganese and the generation of carbon dioxide necessary to attain saturation with respect to manganese carbonate.

Unfortunately it is equally possible that the carbon isotopic values could have been reset during metamorphic recrystallisation (in line with evidence from the oxygen isotopic data) as a result of fractionation between organic carbon and the carbonate phase. Therefore it is not possible to decide whether fractionation of carbon occurred in response to near-surface diagenetic or deeper regional metamorphic processes on the basis of the present data. A more detailed study of the ore-bed would be required to resolve this problem.

The interpretation of the deposition of the Manganese Ore-bed has important consequences on the setting of the Welsh basin at this time. If concentation of manganese is interpreted as occurring above the sediment-water interface in the basin of deposition, then it is likely that the basin was enclosed. Since coarse grained turbidites occur both above (Manganese Grit) and below (Rhinog Formation) the ore-bed, presumably there was no major change in basin depth at the top of the Rhinog Formation. Some factor must have controlled the abrupt change from predominantly sand-rich turbidites in the Rhinog Formation to finer grained turbidites in the Hafotty Formation. This change may have been caused by fault movement at the basin margins, either resulting in a shifting of sources or a shutting off of a coarse clastic supply.

Mohr (1964) found that geochemically the ore-bed and the "normal" Hafotty shales are identical apart from the enrichment of the former in manganese, which argues against a different source for the shales and the manganese, and against a reduction in sedimentation rate in ore-bed times. However the mechanics of manganese production and precipitation still remain unresolved.

Most manganese deposits are Precambrian in age, when the presence of a more reducing atmosphere resulted in manganese being much more mobile within sedimentary systems. The Hafotty ore-bed is similar to the Gondites (metamorphosed Precambrian ore deposits) of India as described by Roy (1976) but since the Hafotty ore-bed was deposited in the more oxidising environment of the Phanerozoic, interpretation is more difficult.

4.11 : Barmouth Formation. (60-200m thick)

The Barmouth Formation is defined by the presence of thick bedded sandstones above the Hafotty Formation (Matley and Wilson 1946). For instance on the western slopes of Garn [SH 616 165] the thick to very thick bedded Barmouth Formation abruptly overlies siltstones with minor sandstones of the Hafotty Formation. The top of the formation and its contact with the Gamlan Formation is also abrupt. The Barmouth Formation is thickest in the southern part of the Harlech Dome around Barmouth and in a NNE-SSW trending zone in the central part of the Dome, thinning towards the northwest and southeast (Allen & Jackson 1985). Thinner bedded, finer grained sequences are rare in this central zone but appear to thicken towards the southeast (e.g. on Y Garn, see BGS Sheet 135).

The Barmouth Formation is comprised mainly of Sand-rich Facies with abundant Amalgamated Coarse Grained Facies. At Garn a nearly complete sequence through the Barmouth Formation is exposed (Figs 4.70. 4.71) containing four cycles (also observed by Allen & Jackson 1985) including coarsening and thickening and fining and thinning trends. However fining and thinning trends are generally better developed, especially at the base of the sequence. Associated with these trends are other vertical changes e.g. in the degree of amalgamation, the type of grading and the occurrence of scour and fill structures. Grading is particularly variable in the coarser parts of the sequence and normal, inverse and multiple grading (divisions S1, 2 and 3) are common. The Barmouth Formation is predominantly comprised of top absent Bouma T_a and T_{ab} units with occasional T_{abc}. Cross-bedding is sometimes present, usually occurring between thick amalgamated sandstone beds e.g. Craig y Penmaen [SH 723 298] and at the same locality rare trace fossils (*Planolites*) are found.
Sole structures indicate that the Barmouth Formation was deposited by currents flowing from the south (Crimes 1970a).

Ripple cross laminae indicate flow in a similar direction but cross-bedding indicates flow towards the west and northwest.

Many of the features described above are discussed further (including examples from the Barmouth Formation) in section 4.5 and 4.6. Thus the Barmouth Formation is similar to the Rhinog Formation in that it was mainly deposited from high density turbidity currents in a highly variable turbidite system. However the Barmouth Formation shows several differences to the Rhinog Formation. In general within the Barmouth Formation the Thin Bedded Facies is less common, mean bed thicknesses are greater and beds are slightly coarser grained (e.g. pebble beds are more common). The Barmouth Formation was therefore deposited in a sand-rich turbidite system with probably a higher sedimentation rate than the Rhinog Formation and thus may have been more proximal or in general received a slightly coarser clastic supply.

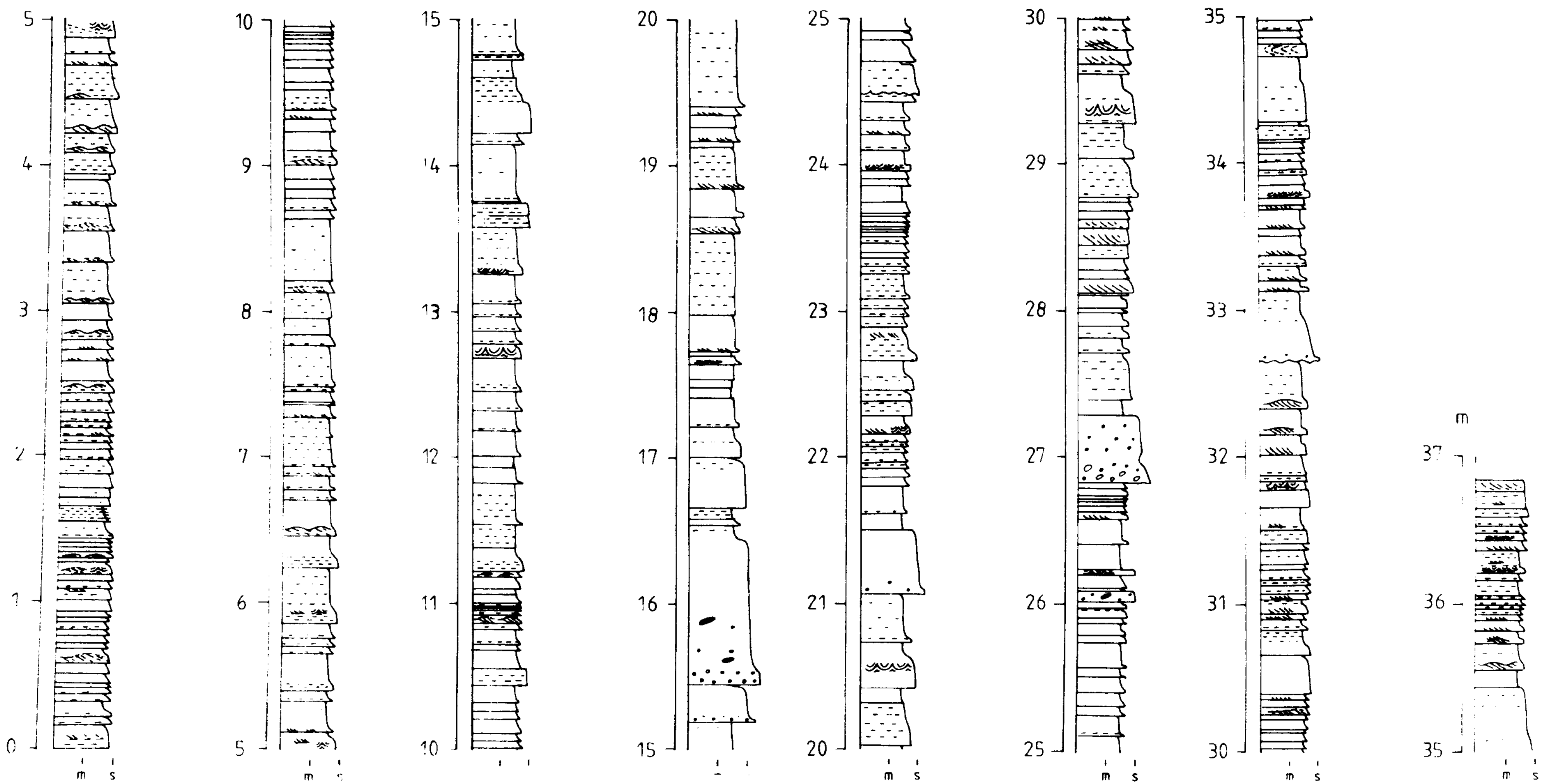
4.12 : Gamlan Formation

The Gamlan Formation is comprised of grey, green (chloritic) and purple siltstones which are locally manganiferous, especially in the upper part of the formation, and some coarse grained greywacke sandstones. The formation outcrops on the northern, eastern and southern sides of the Harlech Dome as well as a small exposure to the south of the Mawddach estuary near Fairbourne (Jones 1933). Its thickness varies from 360m in the north to 230m in the southern part of the Dome (Matley & Wilson 1946), though these estimates include the thickness of igneous sills. The thickness variations are partly due to the presence of thick mappable sandstone horizons, which are thickest and coarsest on the northern and eastern side of the Dome, and are present but thinner in the southeast and northwest. At Moel-y-Gerddi [SH 620315] the differential weathering of the sandstones and siltstones produces a stepped topography, the more resistant sandstones forming low cliffs. At least three amalgamated sandstone units have been mapped, some of which appear to be laterally continuous on the kilometre scale. One such unit- the Cefn Coch Grit, forms the upper few metres of the Gamlan Formation on the eastern side of the Dome, ranging from 2-17m thick and showing thickness changes which suggested to Matley & Wilson (1946), supply from the northeast. The Cefn Coch Grit is relatively widespread, being absent only in the southern and western parts of the Dome. At Barmouth the Grit is replaced by an arenaceous transition between Gamlan Formation and the overlying Clogau Formation (Allen & Jackson 1985). In the Barmouth area coarse arenaceous beds are scattered throughout the Gamlan Formation, but they are thinner than those exposed in the north.

A sequence near the base of the Gamlan Formation was logged near Barmouth shore [SH 617155] (Fig 4.81). The described sequence is typical of the Gamlan Formation as a whole though above this sequence the sandstone and siltstone

FIG 4.81 Barmouth shore log, Gamlan Formation.

Sequence of thin bedded, base absent turbidites.



beds tend to be thinner. The sequence can be assigned to one facies which is composed of beds of eight types, as follows:

1) Coarse grained beds. These have a grain size ranging from coarse sandstone to granule conglomerate. Most beds are about 40-60cm thick, but may range from 7-109cm thick. They are commonly erosively based and are graded including normal distribution grading as well as some coarse tail grading. They may also show repeated grading with thin finer laminae (Fig 4.82). Fig 4.83 shows a type 1 bed (conglomerate) erosively cutting into type 7 beds (graded siltstones), the former containing intraclasts of the latter. Some quartz grains appear to have adhered to the outer surface of one of the intraclasts suggesting that the intraclast was still soft at the time of deposition. This conglomerate is essentially an amalgamated deposit with some variation in grain size both vertically and laterally. Intraclasts concentrate at certain levels suggesting either that they were concentrated at the top of the bed due to their buoyancy relative to the rest of the deposit during flow, or that siltstones were deposited on top and later disrupted by the deposition of the overlying sandstone. Some of the softer intraclasts appear to have deformed under compaction to infill original pore space between grains. The final phase was one of tectonic deformation of intraclasts so that they were partially flattened normal to the cleavage.

Graded conglomeratic or coarse sandstone beds may be overlain by parallel laminated sandstone, which usually only represents a minor proportion of the total thickness of the unit. A unit of a similar type was collected underground at Gwynfynydd Gold Mine [SH 737281] in the Cefn Coch Grit (Fig 4.84), showing distribution grading followed by the full Bouma sequence, forming a 10cm thick unit.

The graded lower division and occasional upper Bouma divisions were deposited in conditions of waning flow, produced by turbidity currents. Top absent Bouma sequences predominate and the presence of repeated grading and thin laminae, possibly of shear origin (the S2 units of Lowe

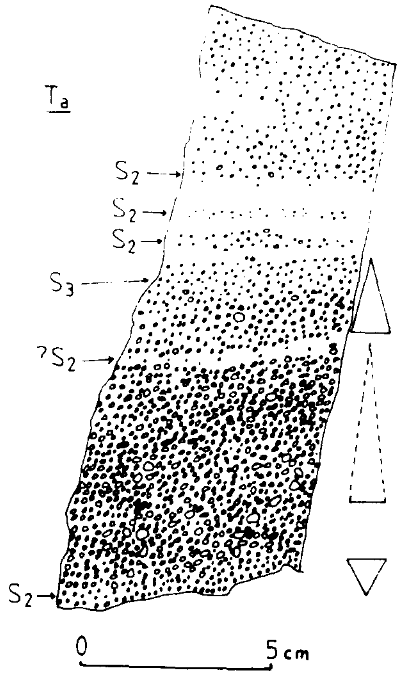


FIG 4.82 Multiple grading in type 1 bed, Gamlan Formation, Barmouth shore.

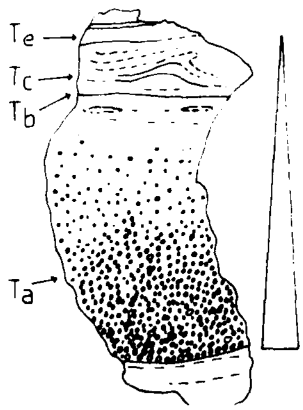
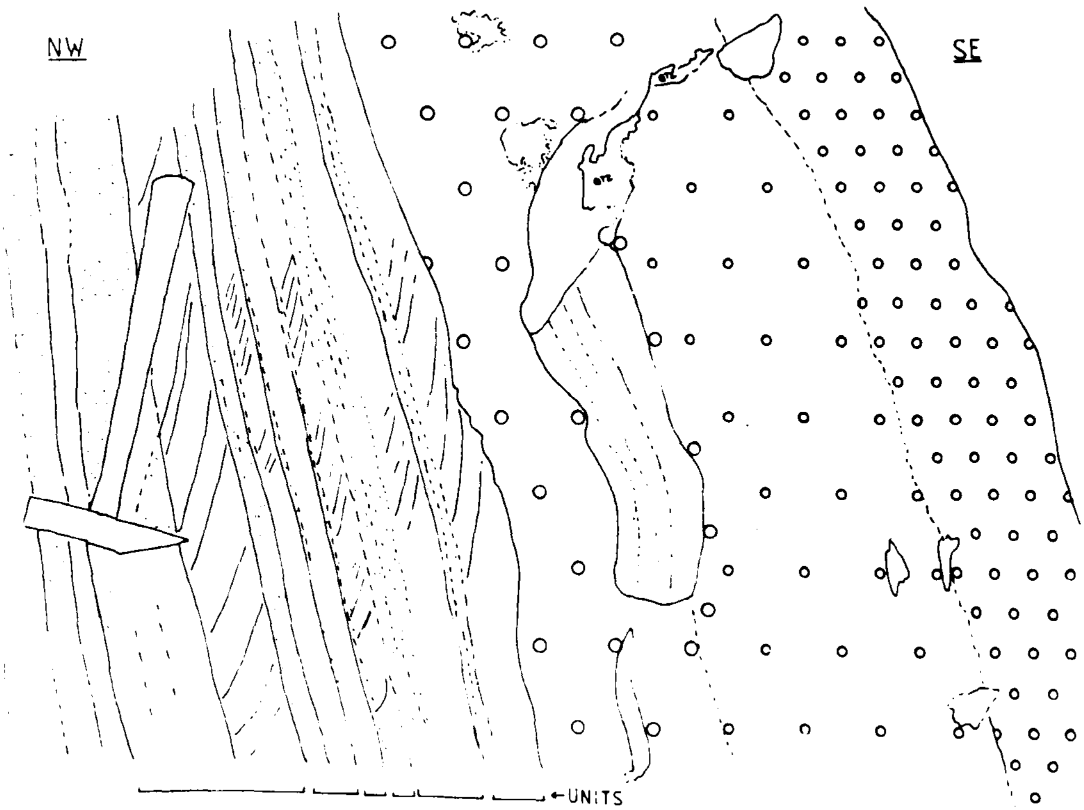


FIG 4.84 Distribution graded type 1 bed, Gamlan Formation, Gwynfynydd.

FIG 4.83 Type 1 bed, Gamlan Formation, Barmouth shore.



1982), suggest that near the base some of these beds may have been deposited from a traction carpet by a high density turbidity current. The presence of repeated S2 units possibly suggests deposition from surging flows with high clast concentrations. S2 units are often followed by suspension deposits (S3), showing simple normal grading.

2) Graded medium to fine grained sandstone beds. Commonly these beds are 5-20cm thick with a maximum thickness of 57cm. They typically lack clear lamination except near the top of beds and show top absent Bouma sequences (largely T_{ab} sequences). They are similar to some of the beds from the Sand-rich Facies in the Rhinog Formation. They are produced by waning flow from relatively high density turbidity currents, though the clasts were never concentrated or coarse enough to produce more than suspension related features (simple grading- S3) near the base of beds.

Type 2 beds also occur in the Cefn Coch Grit in the upper part of the Gamlan Formation. In Afon Gamlan [SH 725243] the Cefn Coch Grit is 8m thick, comprising largely of type 2 beds and rare type 1 beds, consisting largely of T_{ae} and T_{abe} units. The mean bed thickness is about 50cm, with a range of less than 20cm to about 100cm. The occurrence of amalgamated beds, the lack of thin bedded, fine grained beds and the dominance of bed types 1 and 2 suggest that this can be assigned to another facies, similar to Mutti's (1979) facies C1. The occurrence of mainly top absent (though occasionally complete) Bouma sequences and Lowe-type S3 units with rare S2 units suggests relatively high density turbidite deposition.

3) Parallel laminated fine and very fine sandstones passing up gradationally into cross laminated sandstone. They are T_{bc} and T_{bcde}, largely base absent sequences (Plate 4/XXI). Overall the T_{bc} very fine sandstones pass up gradationally into cleaved and laminated siltstones and mudstones. The beds vary in thickness from

PLATE 4/XXI : Faintly laminated graded beds (type 7), Gamlan Formation, Barmouth shore.

PLATE 4/XXII : *Phycodes*, Gamlan Formation, Barmouth shore.

7-38cm with an average of 16cm. The relative proportion of very fine sandstone to siltstone is usually high. This bed type was deposited by relatively dilute turbidity currents compared to unit types 1 and 2, with waning flow from upper plane beds to lower flow regime small scale ripples, in a traction and suspension influenced flow.

4) Ripple cross lamination which passes up gradationally into siltstone and mudstone. Cross lamination occurs as single sets with tangential bases, only rarely forming ripple drift. Beds average 12cm thick, most being 7-15cm thick, though exceptionally they reach 80cm thick. However the rippled part is on average about 2-3cm thick with a range of 1-6cm. Thus the variability in bed thickness is mainly due to variations in the thickness of siltstone in the upper part. A few palaeocurrents were obtained which seemed to indicate flow to the north and west. These beds were also deposited by relatively dilute turbidity currents producing T_{cde} Bouma sequences.

5) Convolute laminated units which pass up into siltstone and mudstone (T_{cde}). Beds average 12cm thick and range from 5-32cm thick. Most beds have truncated tops, for instance in Fig 4.85, which occurs near where the log was measured, in a road cutting 50m east of Barmouth Quarry [SH 617156]. The figure shows a 4cm thick convolute laminated bed, part of an 8cm thick unit (sandstone and siltstone). The laminae overturn in a consistent direction, which if the bed is rotated back to the horizontal, suggests flow towards the west. The bed appears to have originally been cross laminated since truncated cross laminae are still seen, so it was probably originally a T_{cde} sequence. These laminae were either overturned by westward flowing currents or (less likely) as the result of the action of gravity on a westward facing palaeoslope. There was probably differential current drag since the intensity of folding is variable. The fact that there is no mud drape above the overturned convolute laminated sandstone suggests that the

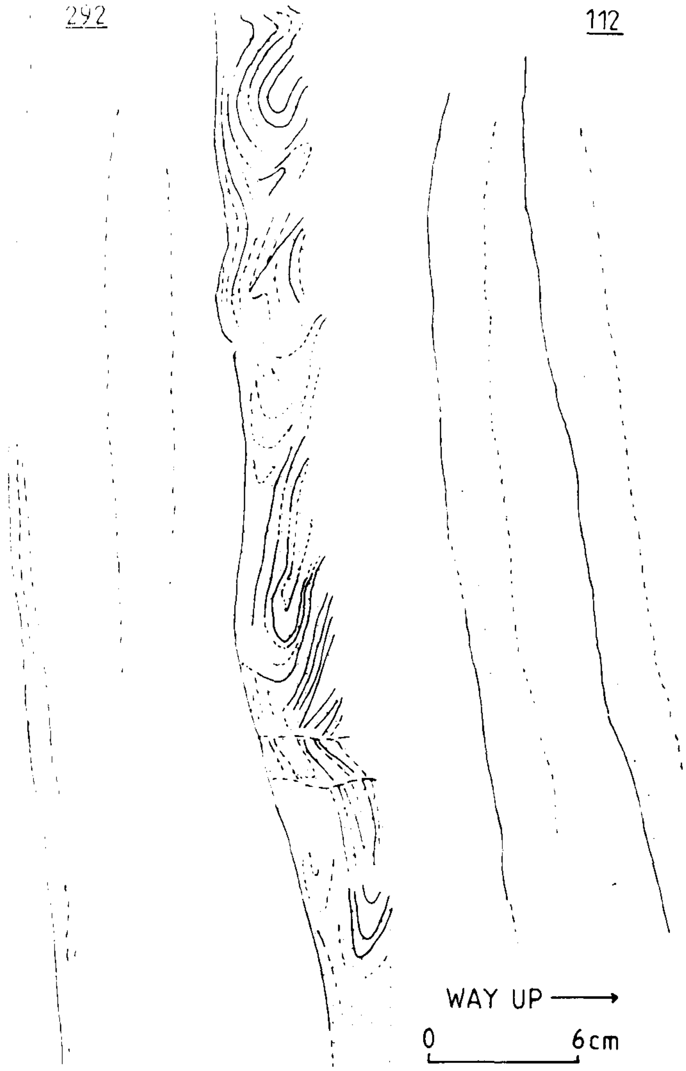


FIG 4.85 Convolute laminated bed, Gamlan Formation, near Barmouth Quarry.

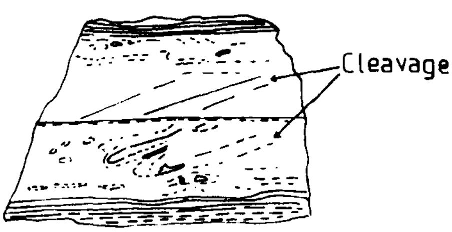


FIG 4.86 Type 7 bed, Gamlan Formation, Barmouth shore.

sandstone bed was deformed and eroded immediately after it was deposited and this event preceded deposition of the overlying siltstone. Deposition, disruption and truncation of the sandstone, followed by siltstone deposition therefore occurred within the same dilute turbidity current event.

6) Thin parallel laminated beds. These beds are similar to bed type 3, except that most of the very fine sandstone beds contain only parallel lamination. They are probably Bouma T_b beds, though many include T_{de} divisions. Beds average 12cm thick, ranging from 2-44cm. The amount of very fine sandstone varies from high in the amalgamated laminated sandstones (T_b) to very low in the thin (2cm thick), parallel laminated units which grade up into cleaved siltstones and mudstones. Thus many of these beds show base absent T_{be} and possible T_{de} Bouma sequences, the product of dilute turbidity currents.

7) Beds with thin, faint laminated, very fine sandstone at the base grading up into cleaved siltstones and mudstones, which are sometimes laminated (Plate 4/XXII, Fig 4.86). The coarser part tends to be lighter and stand proud on the exposure surface. The grading may be readily identifiable or may be indicated by refracted cleavage. Beds normally average 7cm thick and range from 2-19cm, though exceptionally they reach 63cm thick. This bed type was deposited by relatively dilute turbidity currents. Most beds are graded but should not be assigned to T_a of the Bouma sequence (Walker 1967, 1970); instead they are more typical of a silt/mud Piper-type (1978) sequence.

8) Ungraded siltstones and mudstones with occasional wispy laminations. Average bed thickness is 11cm, ranging from 3-42cm thick and are Bouma T_{de} sequences, probably representing hemipelagic deposition.

Kuenen (1953) and Kopstein (1954) suggested that folds in the Gamlan Formation of Barmouth had a soft sedimentary slump origin (involving all the bed types described) with a wavelength of approximately 1m and an amplitude of tens of centimetres (Fig 4.87). The axial traces of these folds are aligned N-S and most of the folds plunge gently towards the south. However the presence of cleavage fans related to the folds and a lack of clear evidence of cleavage overprinting on these folds tends to suggest that the folds were tectonically produced.

Of the eight types of beds described above, type 7 is the most abundant and type 6 is also very common. The prevalence of types 6 and 7 indicate deposition by dilute turbidity currents, with infrequent coarse grained high density turbidites (depositing types 1 and 2). The bimodality of the grain size is therefore related to different depositional processes. Coarser grained sediment was probably transported by different grain support mechanisms in high concentration, high density flows from those which deposited bed types 3-7. The distinct grain size differences in the maximum and mean grain size of bed type 1 relative to bed types 3-7 and the absence of intermediate grain sizes may suggest two possible sources for these turbidites.

Beds are in general laterally continuous on the outcrop scale and erosive bases are rare suggesting a probable tabular sheet-like geometry for most beds. Some beds appear to be clustered within the sequence e.g. the thicker bedded sandstones (bed types 1, 2 and 4).

Palaeocurrents obtained from sole structures in the finer grained, thinner bedded units indicate flow from the south (Fig 4.88) agreeing with Crimes (1970a) e.g. bioturbated flutes east of Barmouth Quarry [SH 618157]. Cross lamination in the same sequence, however, indicates flow towards the west and west-north-west. The difference in palaeocurrents may suggest that turbidity currents flowed in different directions as separate currents or possibly a divergence in current direction in the waning stages of flow. No palaeocurrents were obtained from the coarser

FIG 4.87 Folding in the Gamlan Formation, Barmouth shore.

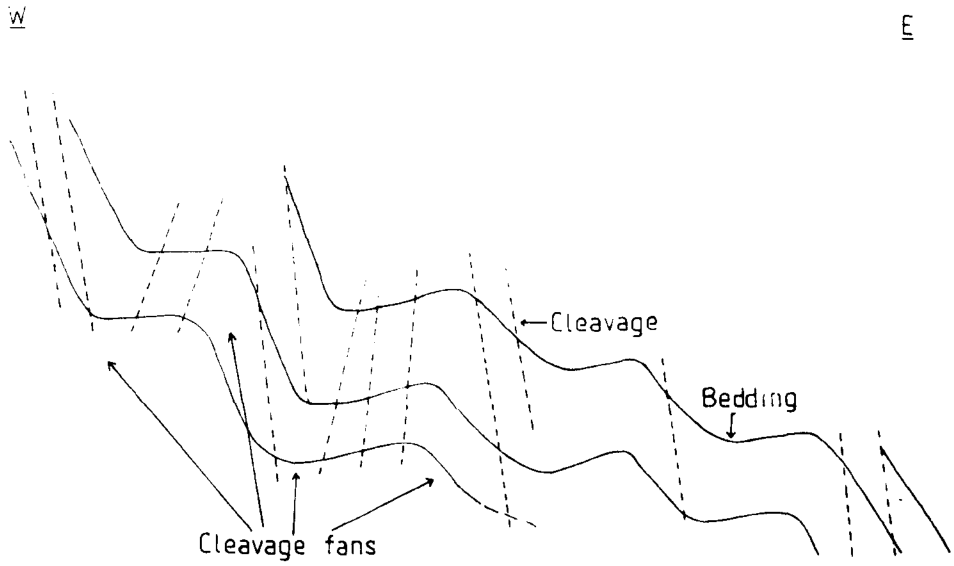
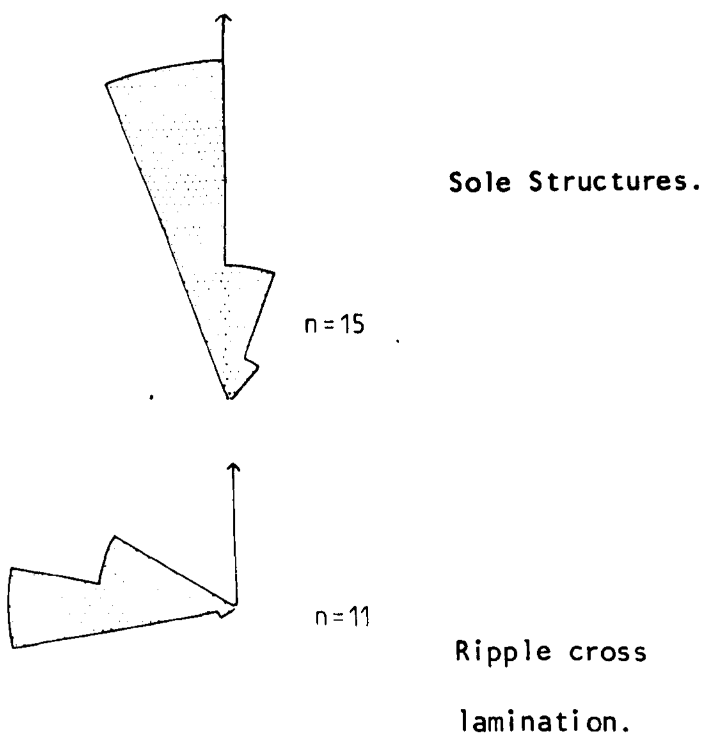


FIG 4.88 Gamlan Formation: Palaeocurrents.



grained beds, but the presence of thicker coarser sandstones such as the Cefn Coch Grit in the north and east possibly suggests a provenance to the northeast. Thus this sequence may represent a temporary return to "proximal"-type (Walker 1967) turbidite deposition derived from the north, interbedded with finer grained more "distal"-type deposition from the south. The external control influencing this switching of turbidites may have been tectonic or eustatic. However the southerly derived turbidites were more frequent and overall volumetrically more important, especially in the southwestern Harlech Dome.

The Gamlan Formation at Barmouth has a single facies which is composed of an interbedding of bed types 1-8. As described above there is some vertical clustering of thicker beds (bed types 1 and 2) and of type 4 beds. However it is not possible to recognise distinct facies composed of particular types of beds as in the facies scheme of Mutti (1979). The Gamlan Formation shows similarities to Mutti and Ricci-Lucchi's (1972) facies D, though bed types occur which one would not expect in this facies. It is not possible to subdivide the Gamlan Formation into facies D1-3 (Mutti 1979).

There is no evidence for coarsening and thickening or fining and thinning upward sequences which might suggest morphological control of sedimentation on a fan. The lateral continuity of beds and the lack of deep scour surfaces suggest a sheet-like geometry and preclude channellisation of these deposits.

The facies is similar to interchannel turbidite deposits which are deposited as channel overbank. Interchannel deposits are typically thin bedded and characterised by base absent Bouma sequences (Mutti and Ricci-Lucchi 1972, Pickering 1980). However the absence of any evidence of channellisation within the sequence argues against this interpretation.

Another possibility based on the prevalence of bed units 4, 6 and 7, is that they may have been deposited in a lobe fringe environment. Coarsening and thickening progradational sequences are considered to characterise the lobe fringe environment on submarine turbidite fans (Walker 1978). Individual beds have a sheet-like geometry but the Gamlan Formation lacks the vertical order within the sequence indicative of lobe progradation. However the weak clustering recognised in the sequence might reflect differences in source or slope processes but since the depositional environment appears to have been generally flat and lacking channels, it is unlikely to have been influenced by the morphology of the depositional surface.

Trace Fossils.

Planolites was found in several parts of the logged sequence. Most are small scale (they average 2mm wide) and rarely exceed 5cm in length. There are some 4mm wide burrows, but they are less common. Many burrows bifurcate. The burrows are filled with very fine sandstone and are often rich in chlorite. At the top of the log a more diverse trace fossil assemblage is found. This comprises of *Phycodes* with spreite (c.f. *Teichichnus*) see Plate 4/XXIII. It is 1cm wide and 55cm long, consisting of a major burrow with 3 offshoot burrows coming from approximately the same point. The spreite may indicate that the producer needed to raise the position of the burrow, possibly in response to sedimentation. Paired vertical burrows were also found (*Arenicolites* and also a possible ?*Diplocraterion*). These trace fossils are similar in many respects to those found in the Rhinog Formation and a similar environmental setting may be envisaged.

4.13 : Conclusions.

The late Precambrian and early Cambrian sequence of the Harlech Dome shows a progressive vertical change from arc/back-arc volcanism (Bryn-teg Volcanic Formation) through fluvial and shallow marine deposits (Dolwen Formation) to turbidites (Llanbedr Formation and above). The early Cambrian succession is therefore largely transgressive, resulting in a gradual deepening of the basin. The turbidite systems of the Harlech Grits Group were either rich in fine sediment (Llanbedr, Hafotty and Gamlan Formations) or sandstone-rich (Rhinog and Barmouth Formations). The Harlech Grits Group is overlain by organic-rich shales (Clogau Formation of the Mawddach Group) above which (in the Upper Cambrian) there was a gradual shallowing upwards trend (Crimes 1970a; Allen *et al.* 1981).

The Dolwen Formation shows changes from alluvial fan through braided and meandering fluvial to shallow water deposits possibly including a tidal shoal facies. Overlying the Dolwen Formation is the Llanbedr Formation which is mainly composed of basinal shales and is a continuation of this deepening upward, transgressive trend.

The common occurrence of thick beds, multiple grading including divisions S1-3 of Lowe (1982), dispersed coarser grains and disorganised sequences indicate that much of the Conglomeratic, Amalgamated Coarse Grained and Sand rich Facies, which dominate the Rhinog and Barmouth Formations were deposited from high density turbidity currents. However cross-bedding in the Sand rich Facies indicates deposition from sustained, low density, probably turbidity current driven, traction dominated flows.

The abundance of thinner bedded, base-absent Bouma sequences in the Thin Bedded Facies (facies D1-3 of Mutti 1979), which dominated sedimentation during the deposition of the Hafotty and Gamlan Formations (Fig 4.89) indicates deposition from relatively dilute turbidity currents which transported mainly fine grained sediment. Rare sandy

FIG 4.89 Summary table of bed types in the Thin Bedded Facies.

The Thin bedded Facies dominates in sequences in the Hafotty and Gamlan Formations and occurs in the Rhinog and Barmouth Formations.

Bed Types.

	Grain Size[1]	Thickness[2] (Common Range, cm)	Bed Classification[3]
1) Laminated mudstones and Siltstones.	Si/M	1-4	T_{de}, $E_{2,3}$
2) Thin graded beds.	vfS-	1-10	E_1, T_3
3) Single set cross laminated beds.	vfS-	1-4	T_{cde}
4) Multiple set cross laminated beds.	vfS-	4-10	T_{cde}, rare $T_{(b)cde}$
5) Convolute laminated beds.	vfS-	4-20	T_{cde}, rare $T_{(b)cde}$
6) Parallel laminated based beds.	fS-	5-20	T_{bcde}, T_{bd}, T_{bc}, T_{be}
7) Graded beds, type 1.	f/mS-	5-20	T_{abcde}, T_{acde}, (T_{abe})
8) Graded beds type 2.	f/mS-	20-60	T_{ade}, T_{ae}, (T_{ae}); S_{1-3}
9) Cross-bedded sets.	c/vcS	6-20	?C_1

[1]Grain Sizes: M= mudstone, Si= siltstone, S= sandstone; vf, f, m, c, vc= very fine, fine, medium, coarse and very coarse.

[2]Thicknesses are given for the sandstone or siltstone part of sandstone, siltstone / mudstone couplets except for bed types 1 and 2.

[3]Bed Classification is as follows:

T_{a-e} from Bouma (1962).
C_1 from Allen (1970).
S_{1-3} from Lowe (1982).
E_{1-3} from Piper (1978).
T_3 from Stow & Shanmugam (1980).

FACIES (Mutti 1979):

Facies D_1, D_2 and D_3 are common.

turbidites in this facies show evidence for higher density flows. This facies may be an interchannel deposit to the coarser grained turbidites or may have a different source (to the east) as a distal lobe to a separate turbidite system. The Thin Bedded Facies was probably preserved during periods when coarse grained clastic supply ceased or switched position, away from the Harlech Dome area and may represent periods of slower sedimentation.

Sole structures indicate that the Rhinog Formation was deposited from flows derived from the north, while cross lamination, cross-bedding and some scour fills indicate palaeocurrents derived mainly from the east (Fig 4.90, Crimes 1970a and present data). Thick bedded, graded and poorly graded sandstone beds dominate in the Rhinog Formation and facies A1, B1 and C1 of Mutti (1979) are common.

In general sand packets are laterally continuous but individual beds tend to be more variable, especially in sections perpendicular to flow (Fig 4.91). Beds are particularly laterally variable in the Amalgamated Coarse Grained Facies and where cross-bedding is common.

Scoured bases are common which may indicate that many beds were deposited on a shallow channelised surface. Deep (greater than 2m deep) scours filled by several beds are absent which suggests that channels tended to be cut and filled within the same overall event. The lateral continuity of beds varies but may be of the order of 2km, so channels would have very high width to depth ratios. This is analogous to fluvial channels which contain low amounts of silt/mud compared to sand; they tend to form very shallow, wide channels with convex upward stratification and are produced by rivers which shift their position during deposition (Schumm 1960, fig 2). In the Rhinog-Barmouth turbidite systems lateral changes in bed thicknesses suggest that a braided pattern of channels may be present. The occurrence of Compensation Cycles indicates that at times the Sand rich Facies was not only channelised but formed

FIG 4.90 Palaeocurrents from the Rhinog Formation.

A) Palaeocurrent rose diagram for sole marks within the areas indicated.

B) Palaeocurrent rose diagrams for:
a) sole marks
b) "wash-outs"
c) cross-bedding.

(from Crimes 1970a)

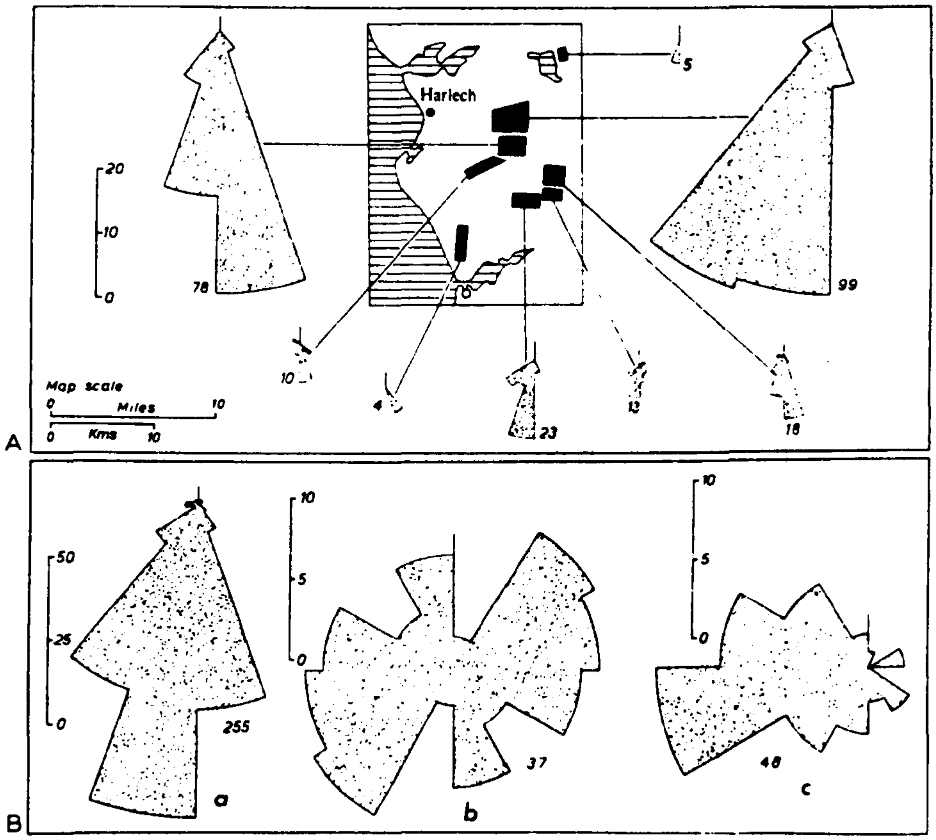
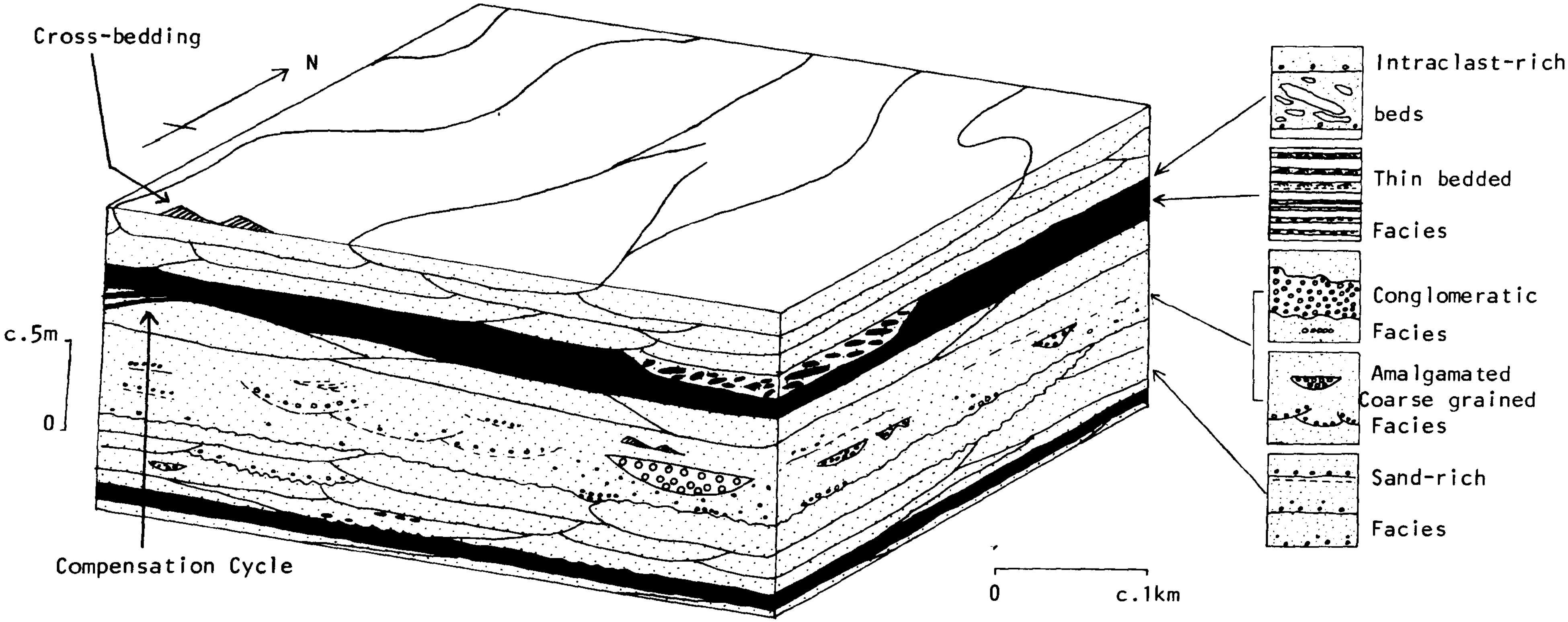


FIG 4.91 Schematic block diagram showing the facies relationships in the Rhinog Formation.

(considerable vertical exaggeration)



upstanding relief (lobes) which progressively aggraded.

The high sandstone-shale ratios (ratios of 1-15 are most common, though several sequences are over 20) and the lack of Thin Bedded Facies indicate that the Rhinog Formation was mainly deposited in a sand-rich turbidite system similar to the sand-rich and very sand-rich turbidite fans described in the literature (e.g. Link & Welton 1982 recorded sand-shale ratios of between 4 and 10 in an Eocene fan from California). The lack of shale in the sequence probably resulted in the formation of a low efficiency turbidite system (c.f. Mutti 1979) where highly mobile turbidity currents may not have been able to develop (Busby-Spera 1985). Instead this type of turbidite system (e.g. the Bullfrog fan, California, Busby-Spera 1985) tends to construct convex-upward profiles with many unleveed channels in its upper part and smooth unchannelled lobes in its lower part. The Rhinog Formation is similar except that smooth lobes are more common and channelisation less common than the Bullfrog fan.

Link & Nilsen (1980) interpret a similar Eocene fan in California as being largely deposited by the abrupt dumping of sand by a series of short-lived distributary channels which may have sometimes extended the length of the fan. Busby-Spera (1985) suggests that this type of fan will tend to pinch out abruptly from turbidite sandstones into starved basin plain shales, so a shift in the position of the sandy turbidite system away from the present area of the Harlech Dome may account for the abrupt upper boundary of the Rhinog Formation. Outer fan lobes tend to form immediately downfan of a channelised midfan composed of laterally coalesced, vertically stacked fans, and inner fan and slope deposits may be difficult to distinguish. However the terms inner, mid and outer fan may not be appropriate within the Rhinog Formation turbidite system where palaeocurrents do not appear to fan outwards and where there is a lack of consistent vertical or lateral organisation (e.g. repeated fining upward sequences). Channelised sandstones also tend to dominate on sand-rich fans though sandy turbidites may not necessarily exit from channels to form smooth lobes. However in the Rhinog Formation it is difficult to define channel and lobe deposits- both had high width to depth ratios and may be difficult to distinguish (c.f. Selley 1985). Gokcen & Kelling (1983) argue that a lack of consistent vertical trends may result from deposition from low sinuosity channels that did not migrate laterally or from channels which were episodically abandoned and reoccupied. This may account for amalgamation in the Rhinog Formation, though the "channels" would have been unstable and have tended to fill hollows on the seafloor to produce generally tabular sand packets.

CHAPTER FIVE

ST TUDWAL'S PENINSULA

CHAPTER 5 : ST TUDWAL'S PENINSULA.

5.1 : Introduction.

St Tudwal's Peninsula occurs at the most southerly point of the Lleyn Peninsula, 25km west of the Harlech Dome. Here Cambrian rocks outcrop in cliff sections on the southern and western shores and are unconformably overlain by Ordovician rocks (Fig 5.1). Precambrian rocks (Gwna Group) outcrop to the west of St Tudwal's (Matley 1928) but the relationship between the Precambrian and the Cambrian in this region is not known.

Cambrian rocks were first proved in this area by the discovery of trilobites in Porth Ceiriad (Ramsay 1881). St Tudwal's was mapped by Nicholas (1915) who established a detailed Cambrian stratigraphy (Fig 5.2):

1) Hell's Mouth Grits (greater than 176m thick- the base is not seen): a sandstone-dominated sequence; mudstones becoming progressively more important upwards.

2) Mulfran Beds (135m thick): manganiferous shales interbedded with occasional sandstones.

3) Cilan Grits (300m thick): a sandstone-dominated sequence.

4) Caered Mudstones and Flags (about 150m thick): a sequence of mudstones, siltstones and fine sandstones.

5) Nant-pig Mudstones (about 66m thick): dark, organic-rich mudstones.

6) Maentwrog Beds (greater than 30m thick- the top is not exposed): a sandstone-dominated sequence near the base, passing upwards into predominantly mudstones.

7) Ffestiniog Beds (greater than 100m thick): siltstones and mudstones.

The succession from the Hell's Mouth Grits to the top of the Caered Mudstones and Flags correlates with the Harlech Grits Group of the Harlech Dome on lithological and palaeontological grounds (see chapter 3).

The structure of the Cambrian of St Tudwal's is

FIG 5.1 Geological map of St Tudwal's Peninsula. (from Nicholas 1915)

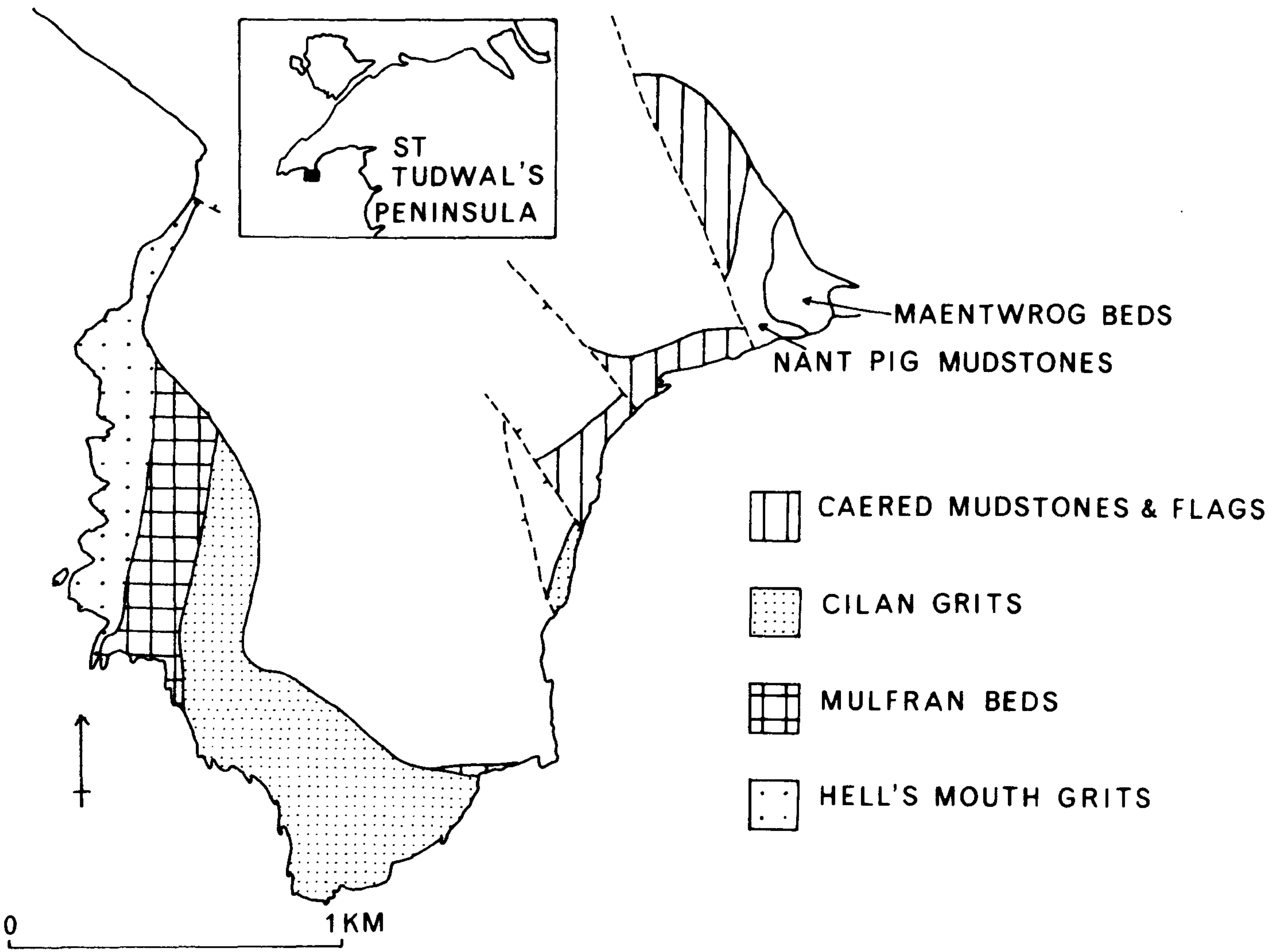
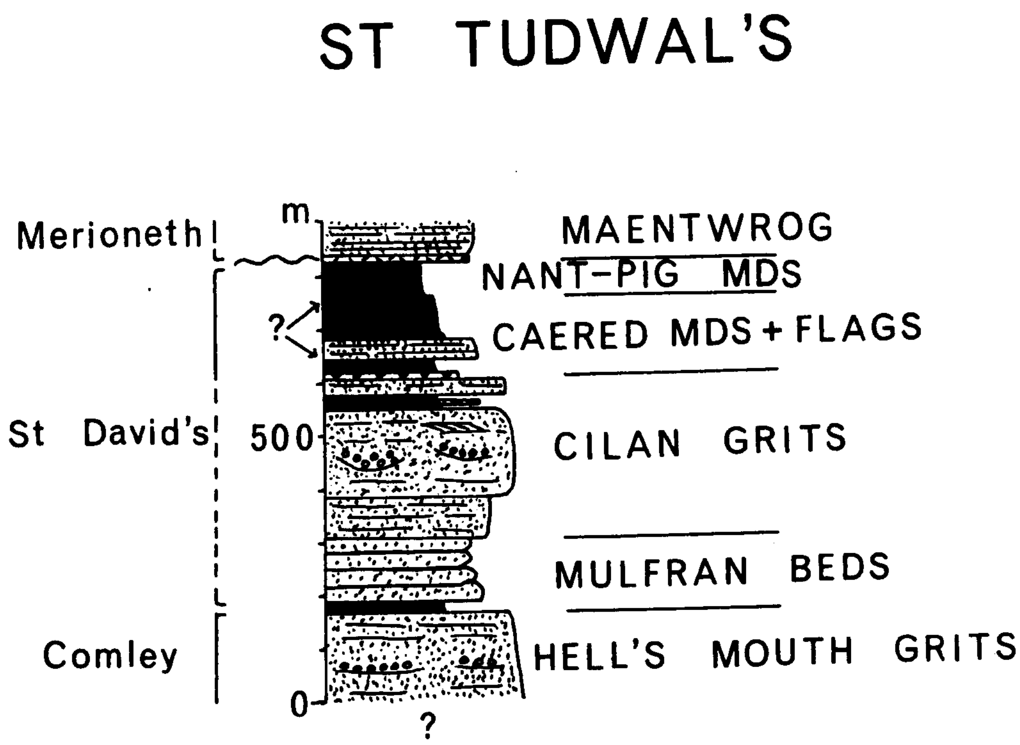


FIG 5.2 The Lower and Middle Cambrian succession on St Tudwal's Peninsula.



comparatively simple. In general Cambrian beds dip at 20 to 40° towards the east and also young in this direction. In contrast the overlying Arenig dips 10 to 20° towards the northeast. A pronounced angular unconformity is seen west of Trwyn Llech-y-Doll [SH 300 234] and in the cliff below Pared Mawr [SH 306 247]. Arenig conglomerates and sandstones progressively overstep younger Cambrian rocks from the Hell's Mouth Grits in the west to the Ffestiniog Beds on St Tudwal's Island East. The Cambrian beds were therefore gently folded prior to the Arenig, probably as a result of tectonic activity in parts of the Welsh Basin at the end of the Tremadoc (Wells 1925; Roberts 1979). The simple outcrop pattern suggested above is complicated slightly by faulting, most of which is post-Arenig. The Cambrian of St Tudwal's is also relatively undeformed and mildly metamorphosed in comparison to the Harlech Dome. Cleavage is more poorly developed on St Tudwal's and metamorphic minerals suggest a lower grade of metamorphism (Roberts & Merriman 1985).

Some of the previously published work on St Tudwal's has concentrated on the sedimentology. Bassett & Walton (1960) were the first to recognise the importance of turbidity currents in depositing the Hell's Mouth Grits, which were derived from the northeast. Crimes (1970a) looked at the sedimentology of the whole succession in order to fit St Tudwal's into a basin-wide model. The same author has also looked in detail at other aspects of the succession: palaeocurrents (Crimes & Sly 1964), the magnetic fabric and its use in determining palaeocurrents (Crimes & Oldershaw 1967), diagenesis (Crimes 1966) and trace fossils (Crimes 1970b, 1975). Field excursions to St Tudwal's have also been reported in the literature (Matley *et al.* 1939; Roberts 1979; Cattermole & Romano 1981).

5.2 : Hell's Mouth Grits.

The Hell's Mouth Grits outcrop on the western side of St Tudwal's peninsula and are the oldest Cambrian rocks exposed in this area. They have a minimum thickness of 176m; their base is not exposed so whether they overlie Arvonian volcanics (Shackleton 1956) or Bryn-teg volcanics is not known, though the St Tudwal's succession generally has more in common with the Harlech Dome than Arfon.

Bassett & Walton (1960) recognised a rhythmic organisation to beds in the Hell's Mouth Grits prior to the formulation of the Bouma (1962) sequence. Palaeocurrents indicate turbidity current flow from the northeast. Crimes (1970a) interpreted these beds as proximal turbidites using the criteria of Walker (1967). Trilobites from the upper part of the sequence were described by Bassett *et al.* (1976) and used to indicate an upper Lower Cambrian age for the Hell's Mouth Grits.

Three main bed types occur within the Hell's Mouth Grits:

1) Thick sandstones of the order of tens of centimetres thick to about 3m.

2) Thin sandstones, ranging from 1 to about 20cm thick, commonly 3-10cm, interbedded with or grading up into type 3 beds.

3) Mudstones and siltstones.

These bed types occur interbedded together, though they tend to be concentrated in certain parts of the sequence. The mean sandstone bed thickness is about 50-60cm in the lower part of the succession and about 70cm in the upper part. The beds range in grain size from mudstone to very coarse sandstone, though occasional granules and large intraclasts also occur. Near the base of graded beds the coarse grains are often clast-supported. Most beds become more matrix rich upwards, the matrix usually being composed of fine sandstone. Beds of fine to medium sandstone with dispersed, coarser grains are common. Sandstone beds range from moderately well sorted to poorly sorted; the latter often have a strong bimodal grain size distribution. For instance at Trwyn Carreg-y-tir (142m), a fine-to-medium sandstone bed is exposed, which contains a population of dispersed "outsize" grains of coarse to very coarse sand grade.

Sedimentary structures.

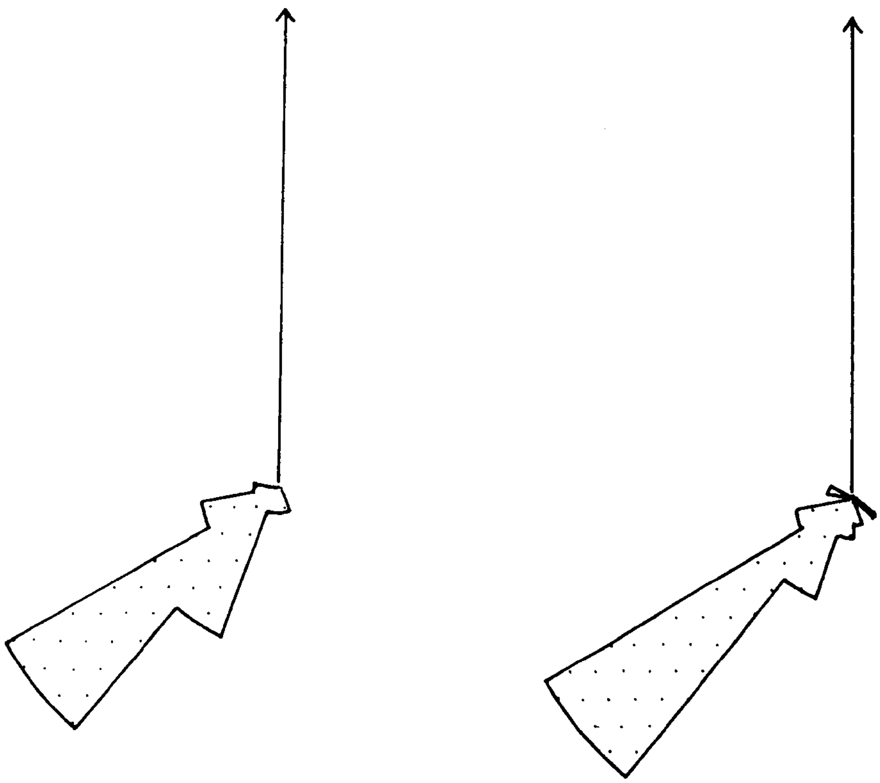
Sole structures are particularly common in the Hell's Mouth Grits; including flute marks, scours, primary current lineation as well as load casts and load affected flow markings. Flow produced erosional sole structures predominate over tool markings and indicate flow from the northeast (Fig 5.3). A range of different types of flow structures occur:

a) Flute casts. They may occur as isolated structures or concentrated in groups and are often associated with larger scale scours. For instance at Trwyn y Fossle (at 32m on Log 1, Fig 5.4) large scours approximately 50cm wide and 10-15cm deep occur with primary current lineation and with grooves and flutes superimposed. The flutes are bulbous, up to 20cm wide and are concentrated in the deepest parts of the scour.

At Trwyn Melyn spectacular sole structures are preserved, in particular on one bedding plane at 23m on Log 3, Fig 5.4 (Plate 5/I). Here there are a series of linear, parallel sided scours which on average are 30cm wide and between 2-8cm deep. The scours contain some low relief flutes and grooves but the larger flute casts are concentrated in the interscour areas. The interscours are also on average 30cm wide; the flutes are about 30cm long and 10cm wide and infilled with a coarse sandstone lag. Small grooves appear to diverge downstream from the flutes. The differentiation of scours and flute-rich interscours indicates regions of greater and lesser scour within the turbidity current. The regular spacing of the scours either resulted from local eddy effects produced once one scour had

FIG 5.3 Hell's Mouth Grits: Palaeocurrents (sole structures).

Bassett and Walton (1960): own data:



n = 444 n = 88

FIG 5.4 Hell's Mouth Grits: Comparison of Logs.

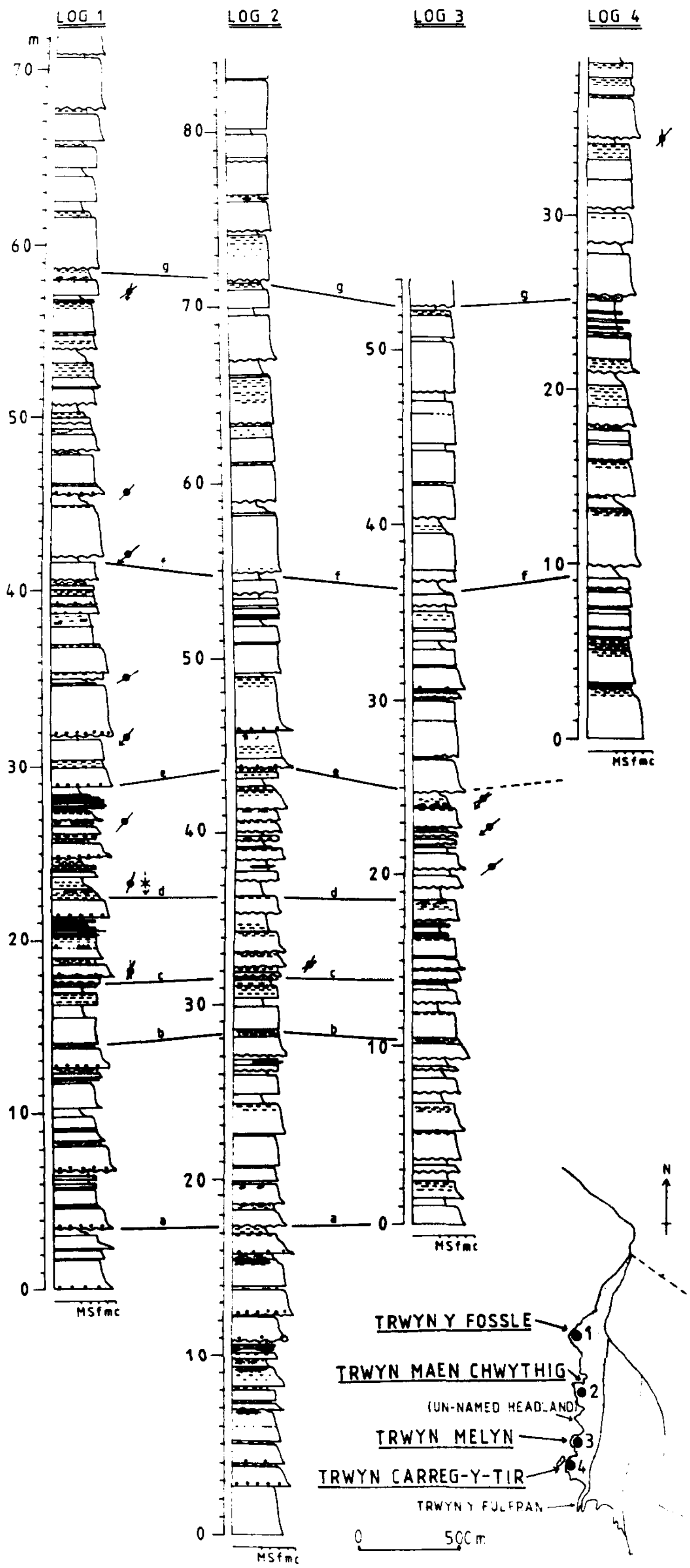


PLATE 5/I : Large flute casts, Hell's Mouth Grits, Trwyn Melyn.

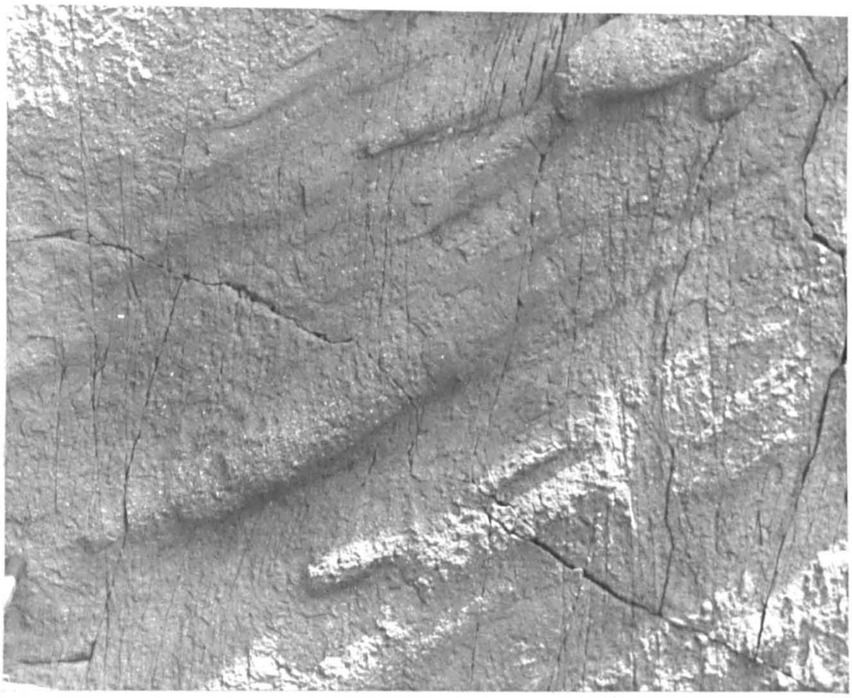


PLATE 5/II : Gutter casts, Hell's Mouth Grits, Trwyn y Fossle.

been initiated or more likely resulted from the action of lobes and clefts in the head of the turbidity current (Allen 1971).

A very large flute (about 3m long and 20cm deep) occurs at Trwyn Melyn. This structure has grooves associated with it which converge on its upstream side. Presumably this current convergence resulted from local eddies, some of which may have been produced by the flute itself.

b) Scours occur in various shapes and sizes and may occasionally contain a coarse fill, indicating traction at the base of a turbidity current. Scours may be irregular or may have parallel sides and be more regularly distributed.

Bassett & Walton (1960) described some steep-sided scours which they called "gutter casts". For instance at Trwyn y Fossle (17m, Log 1) "gutter casts" occur which are 15cm deep and 40cm wide (Plate 5/II). Most have a U-shaped cross section (concave-up), but others are more irregular and have a stepped topography. The "gutter casts" cut into parallel laminated fine sandstones below, truncating and slightly downloading these laminae. These erosional features are filled with medium sandstone, which is parallel laminated near the top (Bouma sequence T_{ab}). Therefore the "gutter casts" are erosional in origin, though the downloading of adjacent laminated beds and the presence of flame structures, some overturned towards the south-west, indicate that they were also load affected.

Some scours contain low angle cross lamination near the base of sandstone beds as part of the scour fill, overlain by unlaminated to poorly laminated T_a sandstone e.g. Trwyn y Fossle (27m, Log 1). In this example the cross laminae dip towards the north, i.e. with reverse sense to the palaeocurrents derived from the flutes. This suggests traction at the base of the flow and reverse currents possibly produced by flow separation within the scour.

c) Load Casts are common and often loading has accentuated pre-existing flow casts, resulting in mud injection on the upstream side of flutes. Load casts and associated flame structures may have amplitudes of up to 40cm.

Graded bedding is also common in the Hell's Mouth Grits, most beds being normally graded. This occurs in two forms:

i) Overall bed grading with the most pronounced grain size reduction occurring abruptly or gradationally near the top of the turbidite unit. Grading is often associated with the transition from T_a to T_{bcd} and since in many beds T_a forms a large proportion of the total thickness of the unit then most grading occurs near the top.

ii) Basal grading is quite common. Coarse tail and distribution grading both occur. Often the larger grains are most prominently graded and there is a slight alignment of the clasts parallel to the current. Many beds are largely clast-supported at the base but the amount of matrix commonly increases upwards. At Trwyn Maen Chwythig (15m, Log 2), for example, superb distribution grading is seen from very coarse sandstone at the base up to coarse sandstone 6cm from the base and coarse-to-medium sandstone above 15cm. In detail however there is slight inverse grading in the basal 1cm of the bed. This may be a high density tractional affect, the result of shearing at the base of the turbidity flow. However multiple grading and other structures indicative of high density turbidity currents are generally lacking, in contrast to the Rhinog Formation of the Harlech Dome.

Other features of the Hell's Mouth Grits:

a) Intraclasts are sometimes present. Most are rounded, though occasional angular intraclasts indicate penecontemporaneous intraformational erosion and deposition. They are rarely greater than a few centimetres long, though they can be up to 30cm long and usually occur near the top of thick sandstone beds, more typical of more distal intraclast deposition (Mutti & Nilsen 1981).

b) Scour fills (see above). Small scours also occur in the finer grained, thinner beds e.g. Trwyn y Fossle (58m, Log 1) where the scours are 3-4cm wide and 0.5-1cm deep. Micaceous siltstone drapes these surfaces and passes up gradationally into parallel lamination though laterally there is a planar truncation surface between the drapes and the parallel flat laminae. These were probably produced within T_b type flow conditions and represent infill of bed topography.

c) Ripples. Small scale ripples are the most common ripple type and occur as T_c divisions, usually as part of complete or more rarely within base absent Bouma sequences. They typically have an amplitude of 1-2cm and have low foreset dips, usually towards the south, though the distribution of palaeocurrents from cross laminae is very variable (Bassett & Walton 1960). Larger scale cross-bedding also occurs rarely e.g. Trwyn Maen Chwythig (41m, Log 2) where the bedforms have an amplitude of 8cm and a wavelength of 30cm. At Trwyn y Fossle (30m) fine grained, cross-laminated sandstone with a set thickness of 5cm is deformed with folded limbs overturned towards the south. This seems to result from syndepositional current shear since the cross laminae appears to infill the depressions between the crests.

d) Convolute lamination occurs occasionally, e.g. at Trwyn Maen Chwythig (9m, Log 2) where the crests have an amplitude of 5cm and a wavelength of approximately 7cm.

e) Sandstone dykes. The tops of some sandstone beds are irregular e.g. Trwyn y Fossle (27m, Log 1) and some have clastic dykes emanating from them, injected into the siltstone or mudstone above. The sedimentary intrusions are rarely greater than 1cm wide though they may form dense, irregular networks, suggesting that the sandstone bed was probably liquefied. The dykes are often concentrated below the down-loaded bases of thick sandstone beds, which possibly caused overpressuring and liquefaction of the sandstone bed below. Structures such as convolute lamination, load casts and sandstone dykes are generally indicative of relatively high sedimentation rates.

f) ?Fracture cleavage. At Trwyn y Fossle (13m, Log 1) a parallel laminated bed contains a near vertical fabric with a spacing of about 1cm. This foliation may be continuous through the whole thickness of the bed or may coalesce. The bed showing fracture cleavage is laterally continuous and can be traced from Trwyn y Fossle to Trwyn Melyn. This may be a tectonic feature, though it appears to be similar to dewatering structures described by Lowe & LoPiccolo (1974). Other faint dewatering structures may be present including possible pipe structures at Trwyn y Fossle (1m, Log 1).

In summary the Hell's Mouth Grits are dominated by T_a-based Bouma sequences, though T_b is also common. Complete Bouma sequences are not uncommon and most beds show a transition upwards into finer grained sediment (T_d). Amalgamated sandstone beds, although present in the Hell's Mouth Grits are not as common as in their lateral equivalent in the Harlech Dome- the Rhinog Formation.

A range of different mudstone and siltstone types are present in the Hell's Mouth Grits interbedded with or grading up from thick bedded sandstones:

a) Parallel laminated green siltstones and very fine sandstones. Laminae are 1-4mm thick with alternating thicker (lighter coloured) and thinner (darker- more organic-rich) layers. Near the top of the succession on Trwyn Carreg-y-tir some of these beds contain trilobites and have a slight blue-grey manganiferous stain when weathered.

b) Parallel laminated dark grey siltstones. Lighter silt laminae 1-2mm thick with slightly graded tops alternate with 1mm thick dark shale laminae. Some beds show even finer scale laminations.

c) Dark grey very fine siltstones and mudstones. These are more massive and only very faintly laminated. They are often micaceous and poorly cleaved. Some beds are of similar type but show different colours- light grey, green, white and yellow (limonitic); the latter two usually occurring directly below sandstone beds.

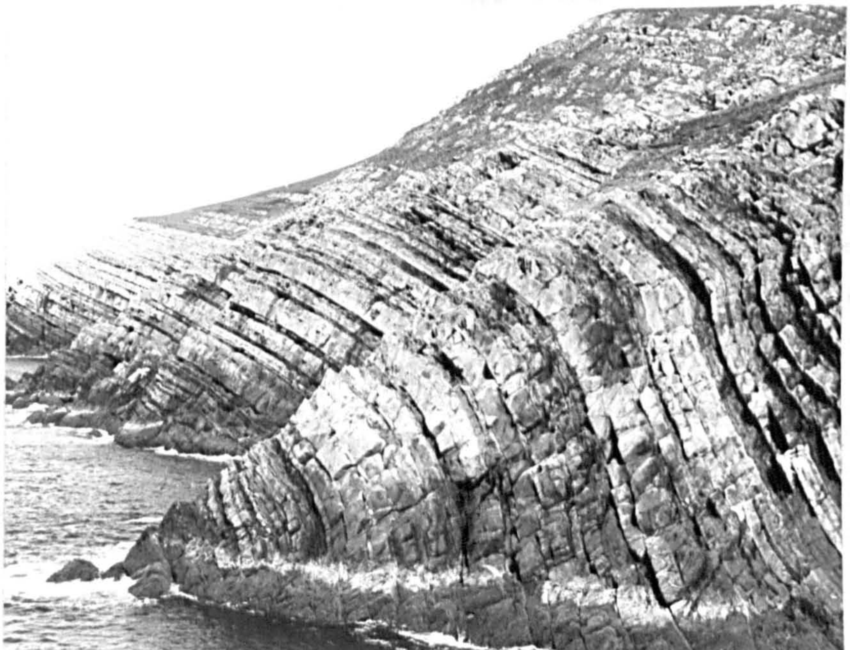
When weathered the mudstones often form 1-10cm scale, ellipsoidal areas bounded by joint planes and commonly give the mudstones a nodular appearance on weathering. Mudstones may occur as relatively thin interbeds between thick sandstones or can form thicker mud-rich sequences of the order of 50cm thick in the lower part of the Hell's Mouth Grits to several metres thick near the top of the Grits on Trwyn Carreg-y-tir. Where thicker mud-rich sequences occur the mudstones are often interbedded with thin bedded sandstones (1-20cm thick, mean c.2-3cm). The sandstones are usually parallel laminated. For instance at Trwyn y Fossle (20m) grey mudstones occur interbedded with 7-8cm thick T_{ab} beds. Many of the sandstones in this facies have a lensoid geometry and have irregular bases and tops. The mudstones are sandy in places and contain intraclasts. Very irregular unlaminated medium sandstone layers are probably clastic intrusions fed from thick sandstones above or below. Cross laminae occasionally occur within the coarser beds in this facies.

Trace Fossils.

Most trace fossils in the Hell's Mouth Grits are hypichnial and exichnial horizontal burrows (i.e. connected and infilled from the overlying sandstone bed). The fill ranges from coarse to fine sandstone, medium sandstone being most common. The burrows are 0.5-1cm wide and on average can be followed for 3-5cm, which is probably a minimum length due to the undulating path of many of the burrows. Most are straight to gently sinuous, though some are more highly sinuous and looped. Several burrows bifurcate (Crimes 1970b-fig 7), sometimes at constant angles and some seem to show a regularity in the spacing of the bifurcations of the burrows. Thus both simple burrows (*Planolites*) and dense clusters of bifurcating burrows (*Phycodes*) are present e.g. Trwyn y Fossle (32m, Log 1), (Plate 5/III). Trace fossils appear to be concentrated in the inter-scour areas of thick sandstone beds and rare examples, which occur at

PLATE 5/III : Bifurcating burrows, Hell's Mouth Grits, Trwyn y Fossle.

PLATE 5/IV : Cliff section, Hell's Mouth Grits, view north of Trwyn Melyn.



the base of scours, usually have subdued relief. Therefore at least some of the burrows appear to pre-date deposition of the overlying sandstone bed. Although common on particular bedding planes trace fossils are generally of low diversity and not very common throughout the sequence as a whole. Burrows are generally most common where finer beds are thickest during periods of slower, finer grained deposition when preservation potential was greater.

Lateral Variation.

It was possible to correlate sections between headlands (Trwyn y Fossle, Maen Chwythig, Melyn and Carreg-y-tir) and thus to compare laterally sequences in the Hell's Mouth Grits (Fig 5.4). Correlation was aided by the tabular nature of the beds, many of which could be correlated directly between logs. Some E-W trending faults make direct comparison at outcrop difficult, but recognition of bed thickness patterns in inaccessible cliff faces enabled correlation with logs (Plate 5/IV). Distinctive sedimentary structures could also be correlated laterally e.g. the "fracture cleavage" bed (bed "b" in Fig 5.4) and the "gutter cast" bed (bed "c").

Bassett & Walton (1960 figs 3 and 5) were able to detect a general decrease in the bed thickness towards the south, between Trwyn y Fossle and Trwyn Carreg-y-tir: a distance of 800m. There is a relatively subtle change though examples can be found which contradict this trend. On the headland scale (c.40-50m) beds may show irregular lateral changes in thickness. In general the thick bedded sandstones have a tabular geometry though there may be lateral changes in sedimentary structures e.g. from T_a to T_b or from T_{abcd} (Trwyn y Fossle, Log 1 58m) to T_{abd} (Trwyn Maen Chwythig, Log 2 71m). Lateral changes from amalgamated to unamalgamated and erosive to planar based sandstone beds are also common as well as lateral variability in grain size. The most pronounced lateral changes result from differences in amalgamation and recognition of individual turbidites within apparently homogenous thick sandstone packets may therefore be difficult. Since few clear consistent trends occur between these sections lateral changes appear to represent irregularity in turbidity current deposition rather than clear downcurrent proximal to distal trends.

The tabular nature of most of the thicker sandstones contrasts with the laterally variable nature of some of the thin bedded sandstones in the mud-rich facies. Some sandstones have irregular thicknesses even on the outcrop scale and seem to fill very broad scours while others show relief on their upper surface (Plate 5/V). Lensoid scours may be concentrated at certain horizons and the variable nature of beds is emphasised by comparing in detail the sequences on three of the headlands (Fig 5.5). The Thin bedded facies is particularly variable e.g at Trwyn y Fossle; 10m, 14m and 18m on Log 1 and its lateral correlatives.

Vertical Sequences.

Fig 5.4 shows two major subfacies which occur above and below bed "e". Below bed "e" sandstone beds are in general less than 1m thick, whereas above beds greater than 1m thick are relatively common. This change is also marked by a reduction in the proportion of Bouma divisions T_b to T_e. This change in subfacies is also obvious at Trwyn y Fossle in the bed thickness diagram (Fig 5.6, at about bed number 40) and diagrams plotting mean bed thickness and sandstone-shale ratio (Fig 5.7). There appear to be thinning upwards trends at 0-25 and 30-50m and a coarsening upwards trend between 50 and 80m. Fig 5.6 however shows that there is considerable variability within the averaged trends in Fig 5.7. Consistent thickening and thinning trends are therefore absent in the Hell's Mouth Grits.

Certain changes can be identified on the larger scale on Trwyn Carreg-y-tir (Figs 5.8, 5.9). Most of the changes appear to be irregular, though the mean bed thickness seems to be greater lower down in the sequence. There is no concommitant overall increase in sandstone-shale ratio, FIG 5.5 Hell's Mouth Grits: detailed comparison of logs.

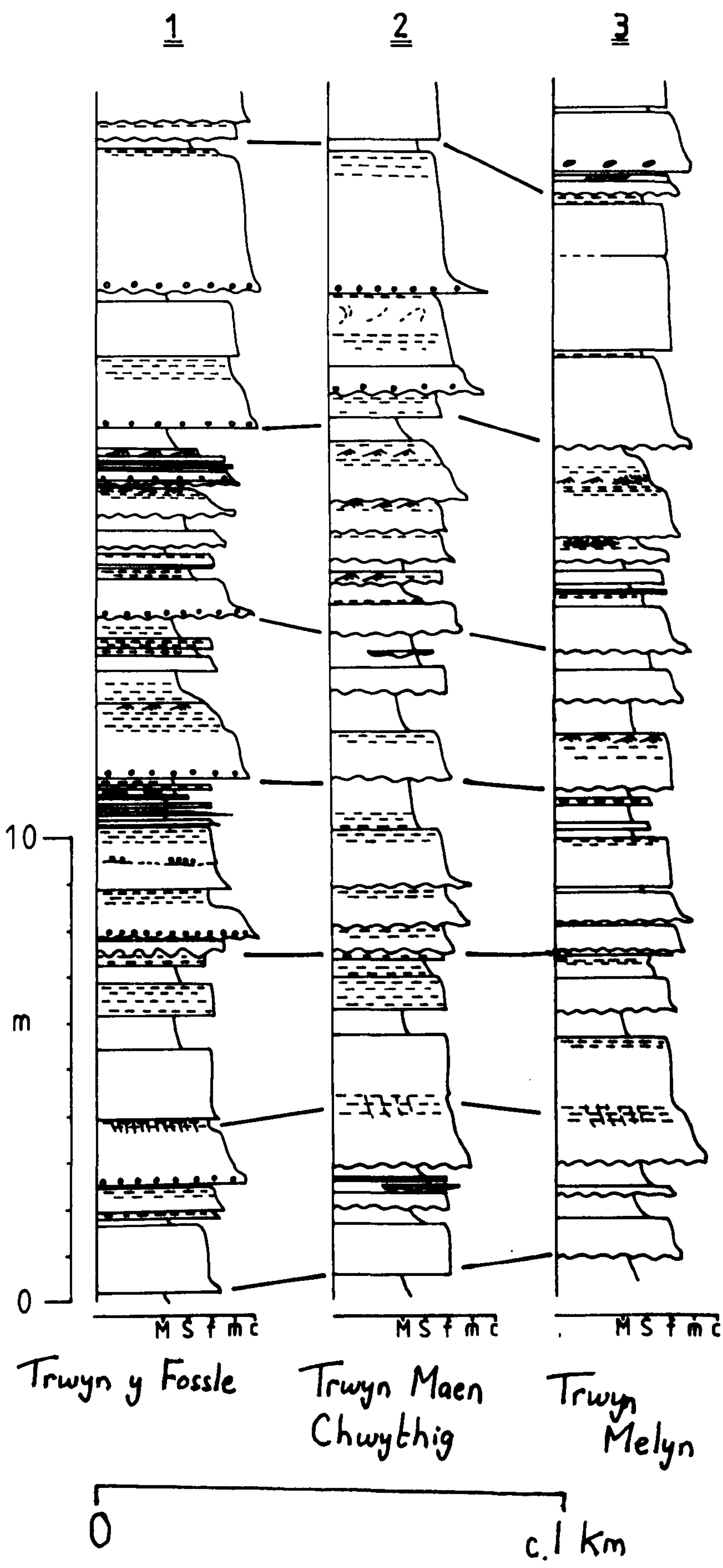


FIG 5.6 Hell's Mouth Grits: a plot of bed number against bed thickness.

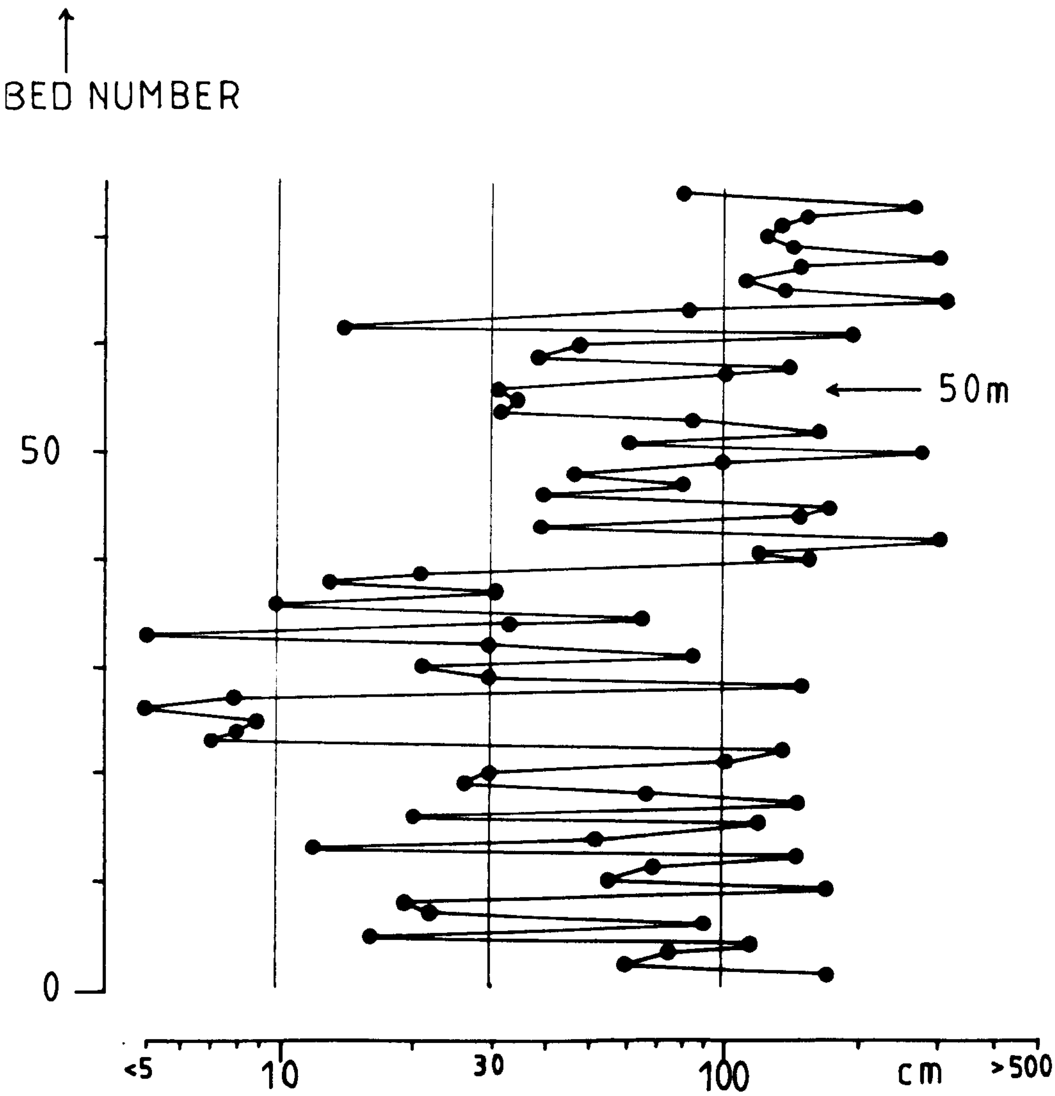
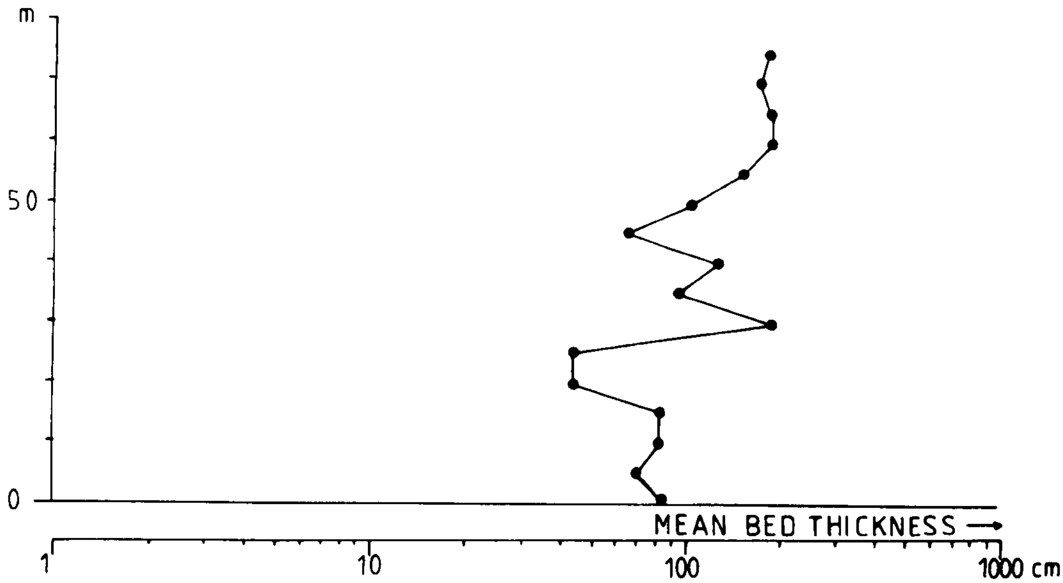


FIG 5.7 Hell's Mouth Grits: Mean bed thickness and Sandstone-shale ratios, Trwyn y Fossle.

MEAN BED THICKNESS (SANDSTONE)



SANDSTONE-SHALE RATIO

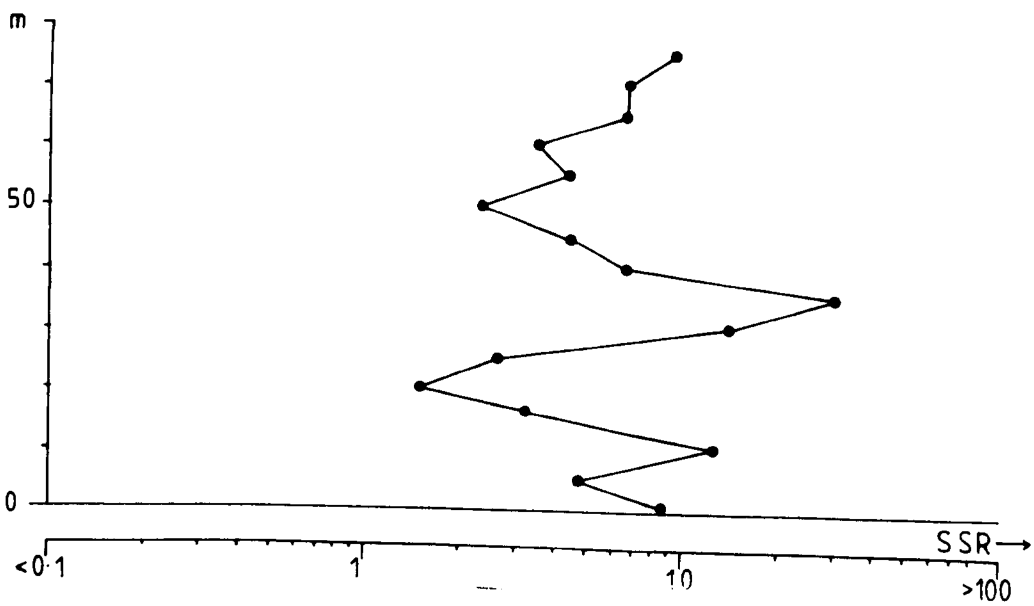


FIG 5.8 Trwyn Carreg-y-tir log, Hell's Mouth Grits.

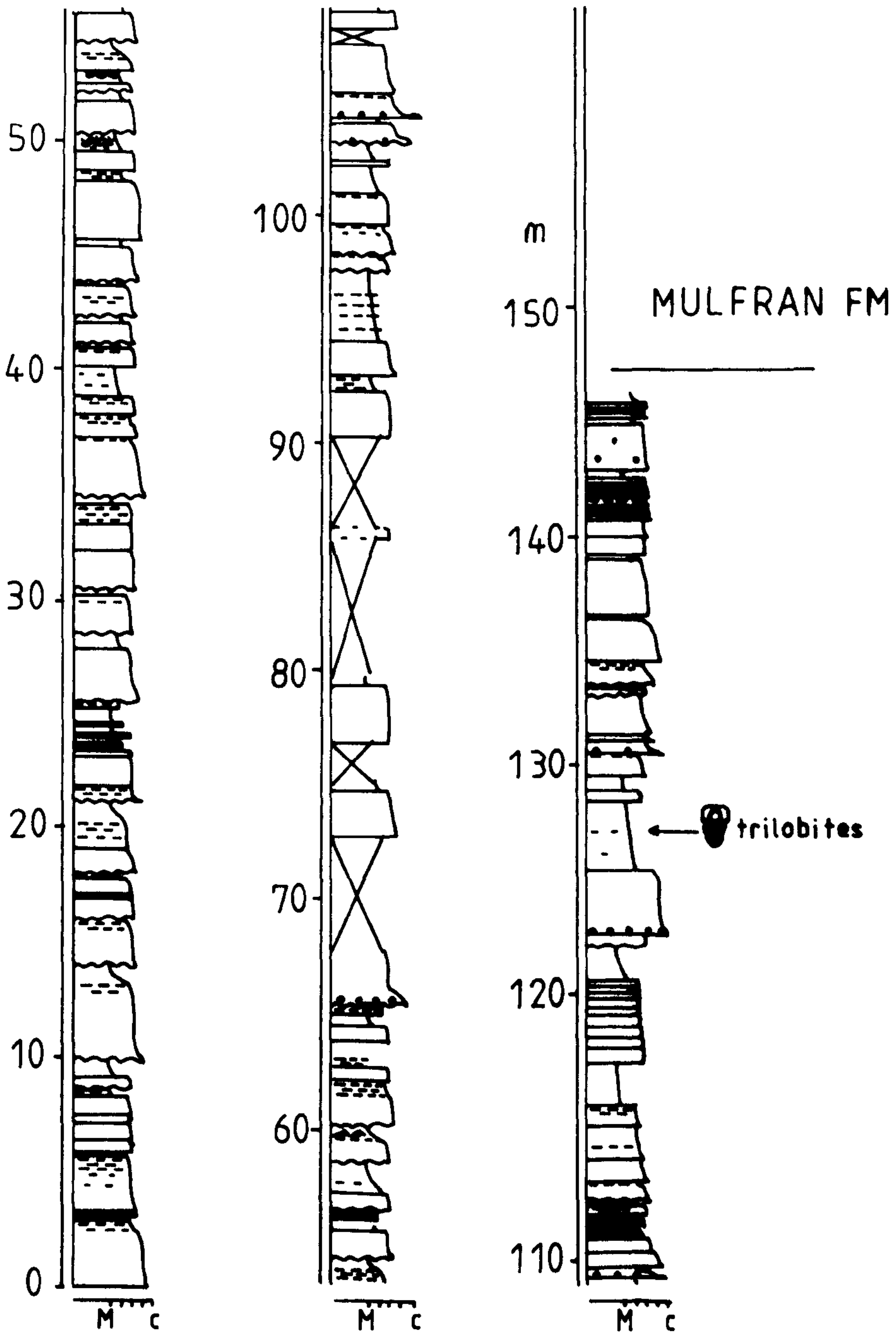


FIG 5.9 Comparison of Mean bed thicknesses and Sandstone-shale ratios at the top of the Hell's Mouth Grits and the Rhinog Formation.

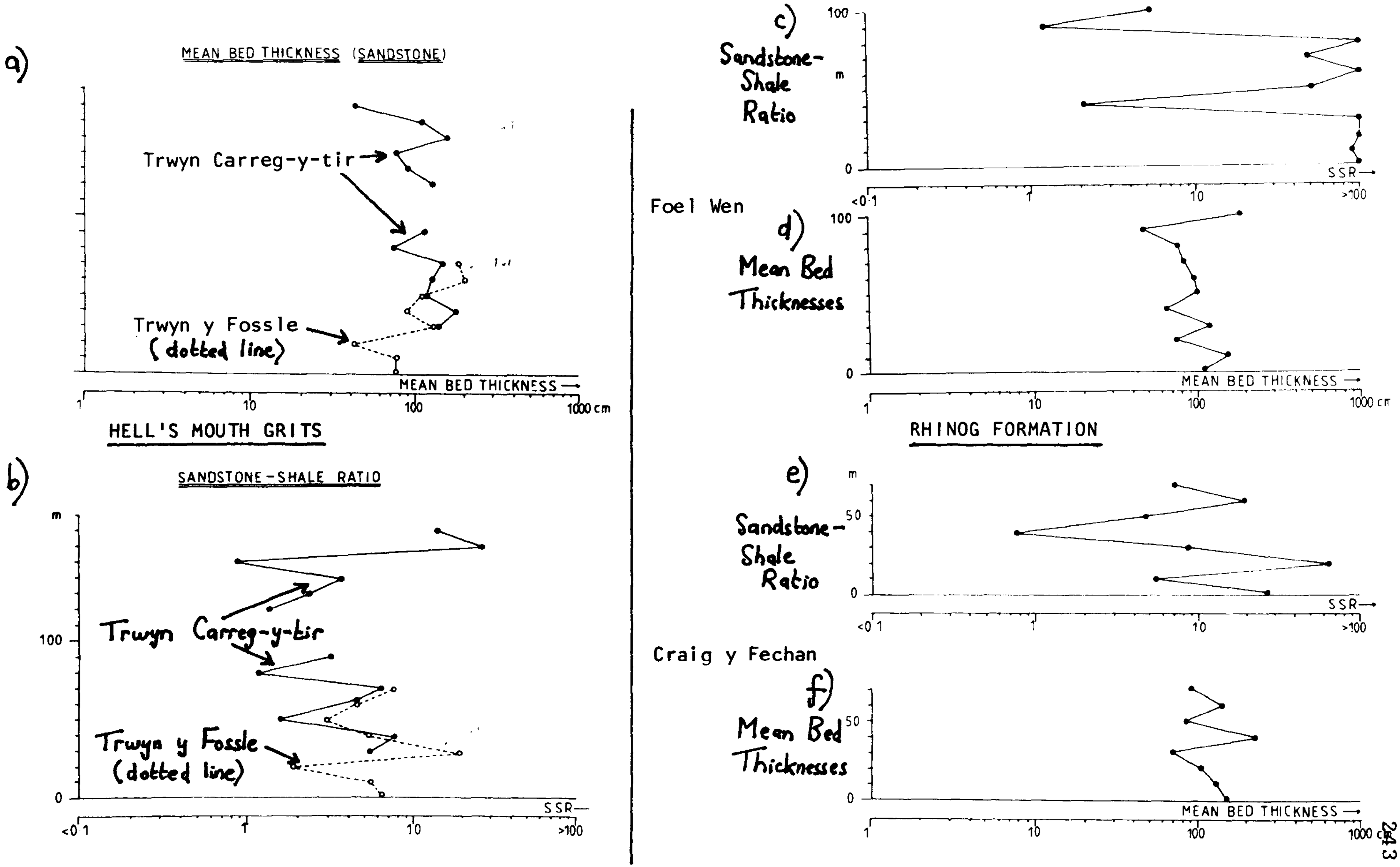
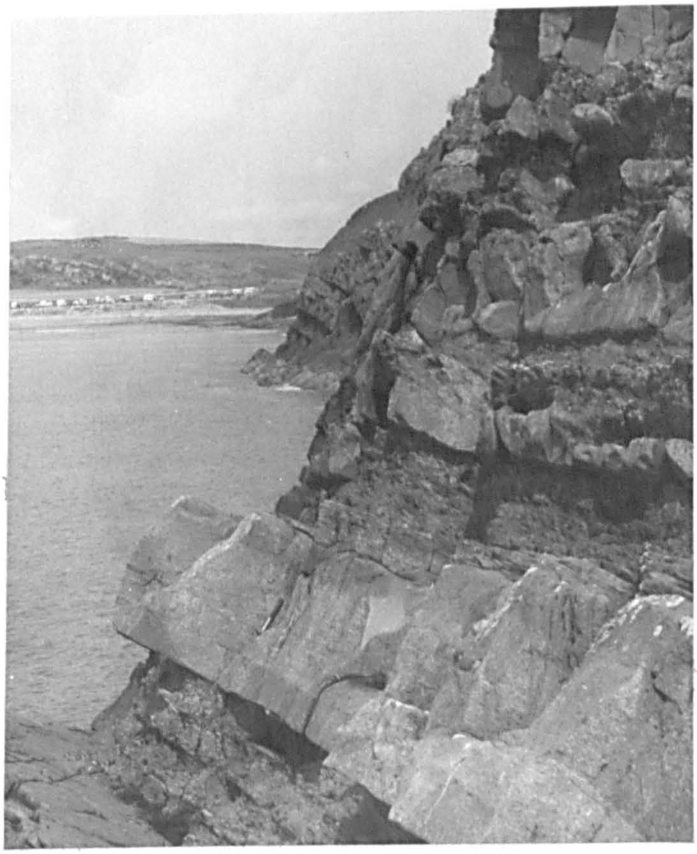


PLATE 5/V : Vertical sequence, Hell's Mouth Grits, Trwyn y Fossle.



though there is a strong bimodality on the smaller scale. The log of Trwyn Carreg-y-tir (Fig 5.8) seems to show increased segregation of sand-rich and mud-rich packages compared to Trwyn y Fossle which occurs lower in the sequence. Thus individual sand packets (of the order of 10m thick) have very high sandstone-shale ratios. However mud-rich packages tend to be thicker in the upper part of the sequence (and probably include many of the non-exposed parts of the succession) and these lower the averaged sandstone-shale ratios. The mud-rich facies contain abundant thin (c. 5cm scale) beds of sandstone.

There is also evidence of increased faunal colonisation higher in the section- trilobites, hexactinellid sponge spicules and an inarticulate brachiopod are described from near the top of the Hell's Mouth Grits (Bassett *et al.* 1976). Siliceous nodules enriched in phosphate occur and "mud-pellet" rock of possible faecal origin may also indicate biogenic activity in the upper mud-rich bands (Bassett & Walton 1960). These trends suggest periods of generally quiet deposition interrupted by brief periods of rapid turbidite sand deposition. There does not appear to be any direct evidence for channelisation of sand within this sequence, though lateral control is limited.

Since the Hell's Mouth Grits are directly overlain by manganiferous beds of the Mulfran Beds the contact between these two units may be regarded as being roughly synchronous with the upper contact of the Rhinog Formation. Correlation between trends in the Hell's Mouth Grits and the Rhinog Formation at Craig y Fechan and Foel Wen (Figs 5.9), however is difficult. This may suggest that:

1) the upper contacts of the Hell's Mouth Grits and the Rhinog Formation are not precisely synchronous, or

2) there were different morphological controls on vertical trends in the Hell's Mouth Grits and Rhinog Formation, or

3) the Hell's Mouth Grits and Rhinog Formation belonged to different turbidite systems.

There must however have been some major ?external control which resulted in a cessation of the supply of sand just before the deposition of the manganiferous deposits.

Conclusions.

The prevalence of thick bedded sandstones containing T_a and T_b sequences indicates that the Hell's Mouth Grits are relatively proximal, sand rich turbidites (in the sense of Walker 1967). These sandstones may be assigned to Mutti's (1979) C1 facies since they show relatively complete Bouma sequences, though T_a forms the largest proportion of most sequences. The lack of direct evidence for channelisation, the lateral continuity of most thick sandstone beds and the lack of obvious, consistent thinning upwards sequences suggests a sheet-like geometry to these beds in a predominantly lobe-type environment. Clear, consistent thickening upward sequences are also lacking and indicate that lobes probably did not prograde and instead aggradation was dominant. Amalgamation, sandstone bed thickness and evidence for faunal activity increase upwards possibly reflecting the development of quieter conditions, only interrupted by brief periods of clustered sand-turbidite deposition.

The thin bedded, mud-rich facies is dominated by more dilute turbidity current activity. The thin sandstone beds are often highly variable in thickness and their relationship with the thicker bedded sandstones is difficult to determine. Since no evidence for channelisation has been found, and thus they cannot be interpreted as channel overbank, the most likely interpretation of the mud-rich facies is that it reflects periods of reduced clastic supply.

The Hell's Mouth Grits contrast markedly with the Rhinog Formation- their lateral equivalents in the Harlech Dome:

1) In general sandstone-shale ratios are lower in the Hell's Mouth Grits.

2) In general average bed thicknesses are lower in the Hell's Mouth Grits.

3) The Rhinog Formation contains coarser grained beds than the Hell's Mouth Grits- very coarse sandstone beds are present but rare in the Hell's Mouth Grits, whereas they are common in the Rhinog Formation.

4) Amalgamated sandstone beds are much more common in the Rhinog Formation.

5) Top absent Bouma sequences are more common in the Rhinog Formation, whereas middle absent sequences are more common in the Hell's Mouth Grits.

6) Sole structures (especially flutes) are more common in the Hell's Mouth Grits, though coarse-filled scours are more common in the Rhinog Formation.

7) Features of high density turbidites (e.g. multiple, inverse grading) are generally rare in the Hell's Mouth Grits but much more common in the Rhinog Formation.

8) The Hell's Mouth Grits seems to show much greater lateral continuity of individual beds than the Rhinog Formation.

9) Evidence for traction current activity (e.g. cross-bedding) is much less common in the Hell's Mouth Grits.

10) The mud-rich facies of the Hell's Mouth Grits is generally more common than the similar Thin Bedded Facies of the Rhinog Formation.

Both the Hell's Mouth Grits and the Rhinog Formation can be classed as proximal turbidites (Walker 1967). Differences between these two units indicate that the Rhinog Formation was deposited from generally higher density turbidity currents than the Hell's Mouth Grits. Whether the Hell's Mouth Grits and the Rhinog Formation were deposited in the same turbidite system is difficult to determine. Both units have relatively abrupt upper contacts with manganiferous deposits suggesting a common external control. The southwestward flowing palaeocurrents in the Hell's Mouth Grits compared to the southward flowing currents in the Rhinog Formation may suggest a fanning of turbidity currents from a point source. However the major differences between the two units seem to suggest that they were deposited in separate sand-rich turbidite systems, though overall the successions are sufficiently similar to suggest that both the Hell's Mouth Grits and the Rhinog Formation were deposited in the same basin.

5.3 : Mulfran Beds.

Near the top of Hell's Mouth Grits the beds fine upwards as part of a 2m thick transition into the overlying manganiferous shales of the Mulfran Beds. The Mulfran Beds are 135m thick and characteristically contain manganiferous mudstones and siltstones which show a distinctive blue-grey weathering colour. Two main facies occur in the Mulfran Beds:

1) Manganiferous mudstone and siltstone dominated sequences, which are particularly common in the basal 15m of the Mulfran Beds. These beds have locally been worked for manganese and Rushton (1974) correlates them with the Ore-bed Shales in the Hafotty Formation of the Harlech Dome. The similarities in succession and lithology indicate that the Mulfran Beds and the Hafotty Formation were probably deposited in the same basin. However the absence of a Harlech Dome-type manganese ore-bed on St Tudwal's indicates:

a) If manganese was precipitated directly from seawater (i.e. the Black Sea model- see Hafotty Formation) then in the St Tudwal's area the deposit was "diluted" by fine grained sediment, indicating a source of sediment in that direction (Woodland 1939). Therefore this hypothesis suggests no ore-bed was formed because the sedimentation rate was higher at St Tudwal's relative to the Harlech Dome.

b) If one accepts the diagenetic enrichment model (Glasby 1974) for the origin of the Manganese Ore-bed in the Harlech Dome then although the sediment was rich in manganese in St Tudwal's, diagenetic concentration of manganese either did not occur within the sediment or did not occur to the same degree as in the Harlech Dome.

2) Interbedded sandstones and manganiferous mudstones. Above the lower sequence of manganese enriched shales, at Trwyn-y-Fulfran [SH 2875 2384], fine sandstone beds form a

larger proportion of the sequence. Unfortunately detailed analysis of this sequence was not possible due to the inaccessible nature of most of the exposures of the Mulfran Beds. However these sandstone beds appear to range from 20-70cm thick, some show poor graded bedding, occasionally with parallel lamination near the top of beds. The sandstone-shale ratio is approximately equal to one. In general thicker beds appear to be clustered within the sequence.

The similarity of the arenaceous facies to parts of the Hell's Mouth Grits suggests that the Mulfran Beds were deposited in similar environments. Both contain abundant, relatively thick bedded turbidites (e.g. T_{abe} sequences), though the Mulfran Beds have a lower sandstone-shale ratio and tend to be thinner bedded than the Hell's Mouth Grits. Thin laminae within the manganiferous shales may indicate more dilute turbidity current activity. The presence of manganese in these mudstones may indicate that manganese was being precipitated along with or adsorbed onto, the background sediment- silt and clay. The thicker development of manganiferous sediments on St Tudwal's and the absence of a true ore-bed may argue for preferential diagenetic enrichment in the Harlech Dome. There was also a greater supply of sand (probably from the north) on St Tudwal's at this time relative to the Harlech Dome.

5.4 : Cilan Grits.

The Cilan Grits are 300m thick and outcrop in the southern part of St Tudwal's Peninsula around Trwyn Cilan and a small fault-bounded exposure on the west side of Porth Ceiriad (Fig 5.1). This unit is dominated by thick bedded graded sandstones in contrast to the overlying Caered Mudstones and Flagstones and lack manganiferous shales in contrast to the underlying Mulfran Beds. The Cilan Grits can be divided into four main parts (Nicholas 1915; Rushton 1974):

1) 65m of fine grained, relatively thin bedded sandstones. This unit fines near the top to green shales and thin bedded sandstones.

2) 163m of coarse, thinly bedded sandstones which thicken upwards. In the upper part of the unit there are coarse massive sandstones which are often amalgamated or separated by thin shale bands. Pebble bands also occur.

3) 38m of red and green mudstones with occasional thin sandstones.

4) 33m of relatively massive sandstones.

Unfortunately, because of the inaccessibility of many of the exposures of the Cilan Grits it was only possible to study the Cilan Grits in detail at two localities:

a) Trwyn Cilan [SH 2940 2305] where sandstone-rich facies in the middle part of unit 2 is exposed.

b) Foreshore, from east of Trwyn Cilan [SH 2967 2310] to west of Trwyn Llech-y-doll [SH 2995 2336], which gives a complete section from the upper part of unit 2, through units 3 and 4 and into the Lower Caered Mudstones.

Trwyn Cilan.

The sequence at Trwyn Cilan is mainly sandstone dominated. The mean sandstone bed thickness for the whole sequence is 63cm, and the mean sandstone-shale ratio is 11. There is, however, a progressive increase in mean bed thickness from 30cm at 0-5m to 83cm at 20-25m and an increase in sandstone-shale ratio from 2 at 5-10m to 19 at 15-20m (Fig 5.10). Two facies can be distinguished:

Facies A.

This facies is dominated by T_a and T_{ab} sandstones about 30-40cm thick. Beds may have planar or erosive bases, are often graded and may contain convolute lamination near the top. Mudstones and siltstones are usually thin, are often interbedded with thin lenses of sandstone but only make up a relatively small proportion of the sequence.

One of the more interesting features of these sandstone is the abundance of cross-bedding. Palaeocurrents from the cross-bedding are highly variable, though there appear to be two modes, one towards the west and northwest, the other towards the south-south-west (Fig 5.11). This contrasts with palaeocurrents derived from sole structures which indicate flow towards the southwest (Crimes 1970a). Usually the cross stratified sandstone beds are lenticular and have a maximum thickness of between 5 and 20cm, which is also the amplitude of the bedforms. At 7m (Fig 5.10) cross-bedding of up to 9cm amplitude and 120cm wavelength, dipping towards the south-south-west overlies a T_{ab} sandstone (Plate 5/VI). The former is composed of coarse sandstone, the latter is medium sandstone with occasional scattered coarse sand grains and there is a relatively sharp boundary between the T_{ab} sandstone and the cross-bedding. However whether the traction current which produced the cross-bedding reworked and winnowed the underlying sediment or supplied its own sediment is difficult to determine. Above this bed, bounded above and below by siltstone, is a lenticular sandstone which reaches a maximum thickness of 18cm. Tabular and trough cross-bedding occur and indicate current flow towards the south-south-west and the west. This bedform also shows different stages of growth from low angle to steep foresets. The abrupt upper and lower contacts indicate that this bed was probably not produced by a simple waning turbidity current but by some type of traction current flowing in a different direction to the main turbidity

FIG 5.10 Trwyn Cilan log, Cilan Grits.

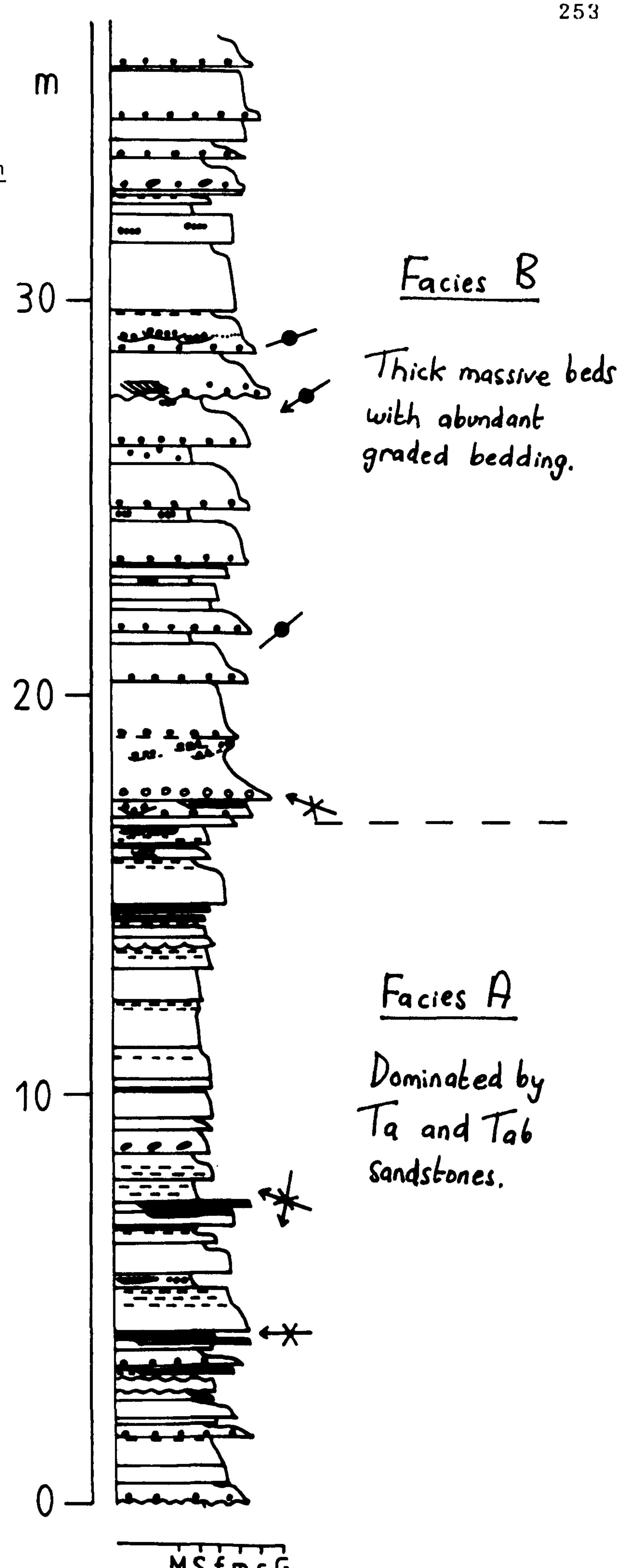
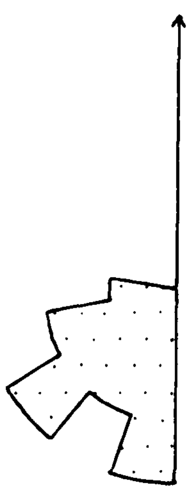


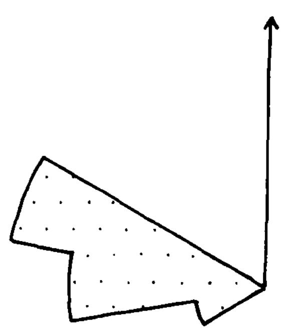
FIG 5.11 Cilan Grits: Palaeocurrents.

sole structures:

cross-bedding:



n=11



n=8

PLATE 5/VI : Cross-bedding, Cilan Grits, Trwyn Cilan.

current flow direction. This traction current was variable in strength and direction and since there is no obvious bed nearby the current could have reworked, it probably supplied its own sediment. The origin of this cross-bedding is probably similar to that described in the Rhinog and Barmouth Formations of the Harlech Dome.

Facies B.

Above 18m there is an abrupt change to thick massive sandstone beds (140-160m . thick) with granule conglomerate near the bases of beds. These beds are graded; normal grading is the most common type though some show reverse grading (S2) near the top of some beds. This facies also contains cross cutting scours which are often lined with coarser sediment. The scours are of the order of 50cm wide and 30cm deep. Loaded bases are also common, sandstone-shale ratios are very high and T_a divisions predominate. These characteristics are very similar to the Amalgamated Coarse Grained Facies of the Rhinog and Barmouth Formations and suggest deposition from high density turbidity currents.

A similar facies also occurs west of the logged section on Trwyn Cilan foreshore and includes several pebble conglomerate lenses.

Lateral Variability. On viewing Trwyn Cilan from the northwest most thick sandstone beds appear to be tabular. However some beds approximately 1-3m thick (which occur directly below the logged sequence in Fig 5.10) seem to be lenticular. These features possibly represent single channel-fill events since the sandstones appear to be massive and thickly bedded. There is no evidence for larger scale channelisation.

Trwyn Cilan to Trwyn Llech-y-doll Foreshore.

Two major facies occur in this section: a sand-rich facies and a mudstone-siltstone facies, the latter of which occurs between 10 and 30m in Fig 5.12 (see also Figs 5.13 and 5.14). There is a transitional sequence between the two facies below (5-10m) and above (30-40m).

FIG 5.12 Trwyn Cilan to Trwyn Llech-y-doll section, Cilan Grits.

G: green
g: grey
R: red

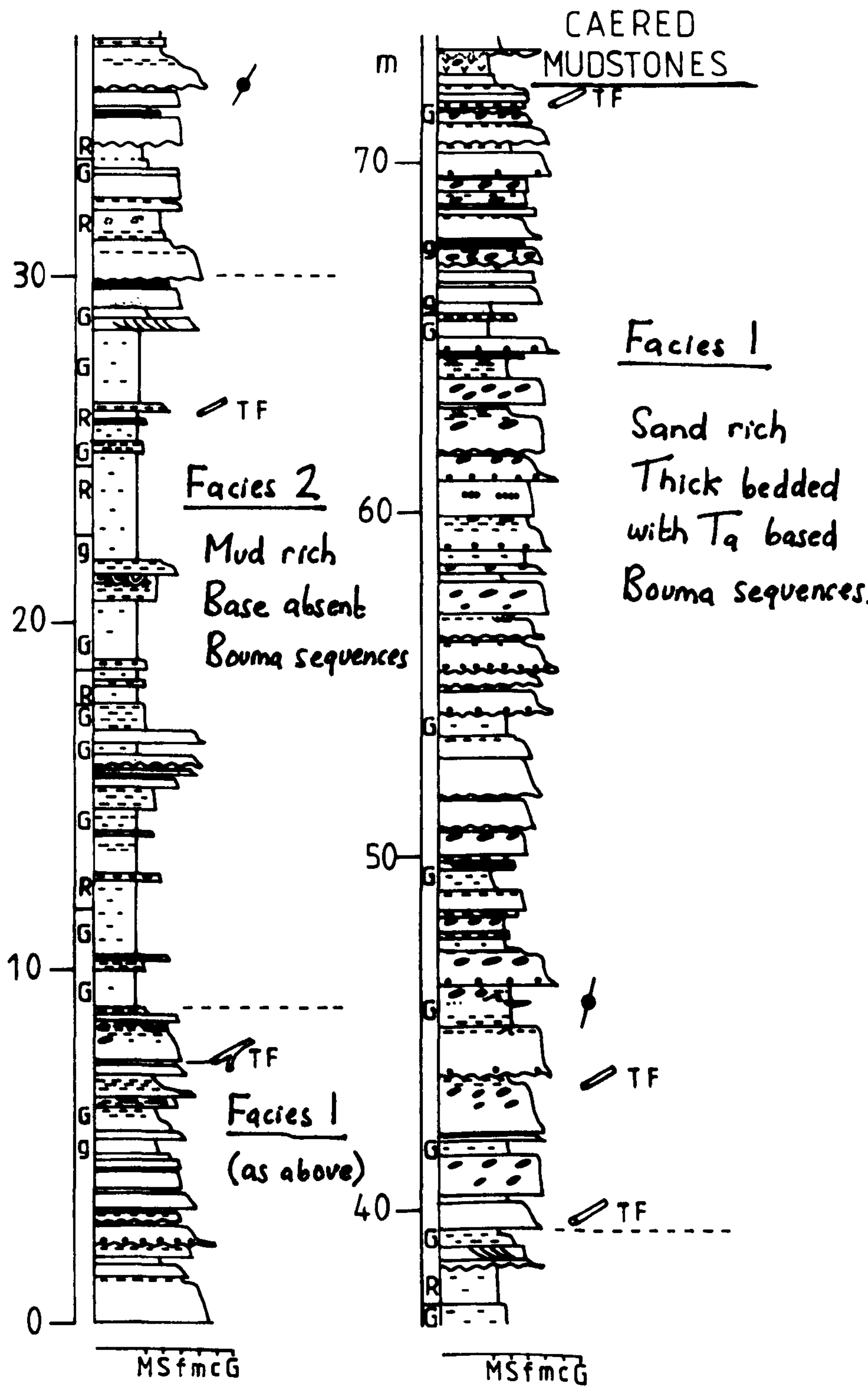


FIG 5.13 Cilan Grits: a plot of bed number against bed thickness.

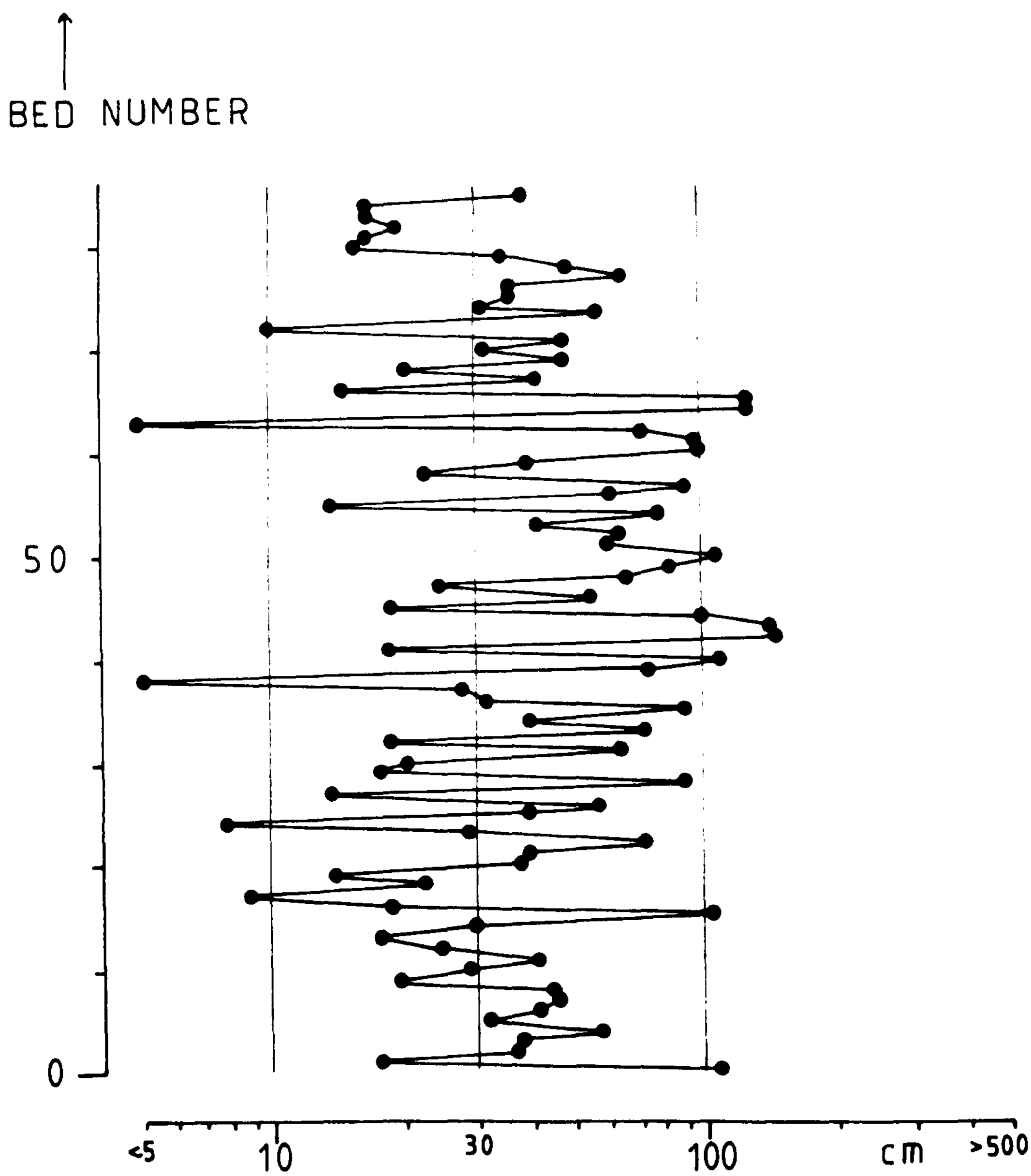
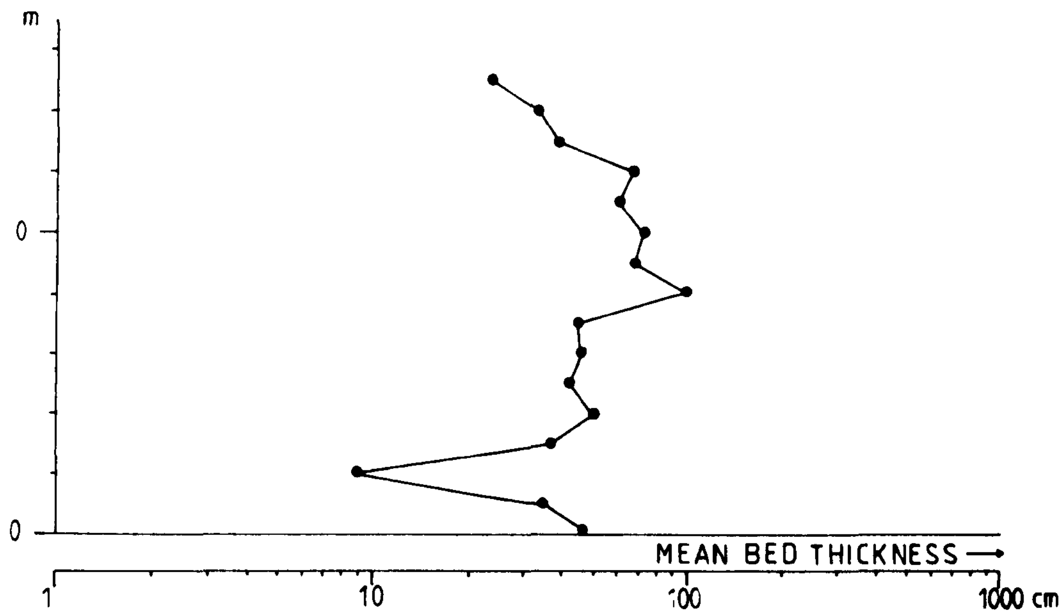
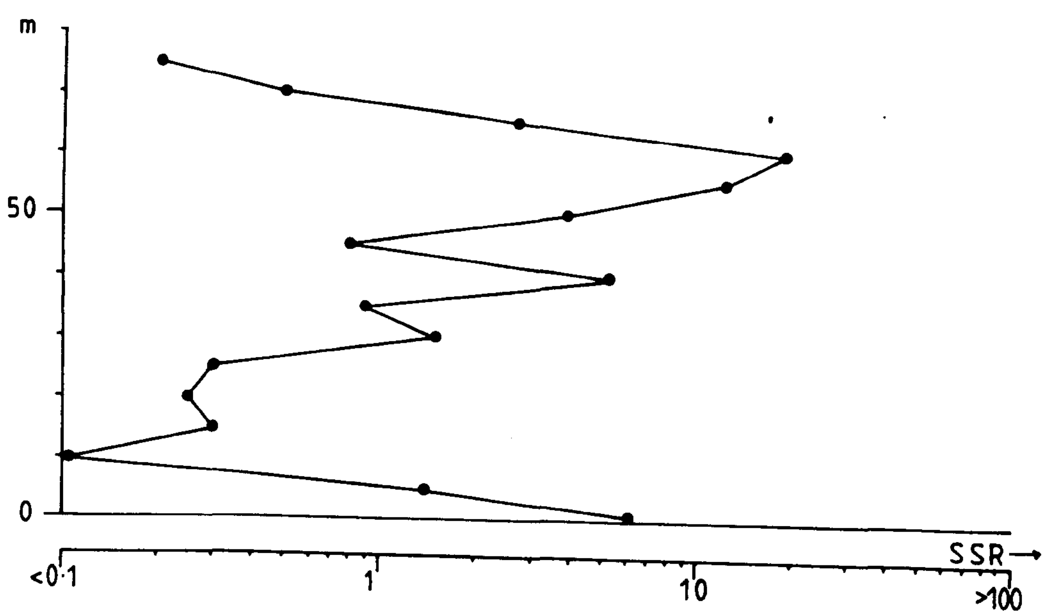


FIG 5.14 Cilan Grits: Mean bed thicknesses and Sandstone-shale ratios.

MEAN BED THICKNESS (SANDSTONE)



SANDSTONE-SHALE RATIO



Facies 1.

This facies is generally sand-rich and contains sandstones with bed thicknesses of up to 2m but a mean of about 40cm. Sandstone-shale ratios are high, averaging about 3. Grain sizes range from mudstones to coarse sandstone. Many sandstone beds are normally graded, the grading usually being restricted to the bases and tops of beds. Distribution and coarse tail grading both occur. Many beds have dispersed coarse grains in a matrix of fine to medium sand so that there is a strong bimodal distribution of grains; a feature often found in thick, massive, poorly graded beds. Clear directional flow-produced sole strucures are rare though irregularly scoured bases are present as well as loaded bases. Intraclasts are common near the tops of many beds, some are angular suggesting nearby penecontemporaneous erosion and deposition. Parallel lamination is common near the top of beds (as part of T_{ab} sequences) and cross lamination is also quite common. Occasionally cross lamination has undergone soft sediment deformation to convolute lamination. Cross-bedding is also present.

Many beds have irregular undulating tops. Some, for instance as shown in Fig 5.15, abrupt, scoured surfaces are present which may be infilled by thin bedded siltstones (as in this case) or by sandstones, which may represent localised channelised flow. At one place a 30cm thick sandstone bed thins to nothing to the west; this occurs within thin bedded siltstones and indicates, at least locally shallow channelised flow (Fig 5.16). However in other places the tops of beds are more chaotic, which may result from liquefaction of the bed on dewatering and sediment injection into overlying siltstones. In other cases a chaotically bedded sandstone occurs downcurrent from a cross-bedded unit which was probably produced by current shear. Fig 5.17, however shows complex patterns of amalgamation and downward truncation of several thin siltstone beds; the former is probably produced by sand remobilisation due to compaction, the latter may be produced by slumping. Thus instabilities resulting from loading, Cilan Grits: Sedimentary structures.

(scale about 5cm represents 1m)

FIG 5.15

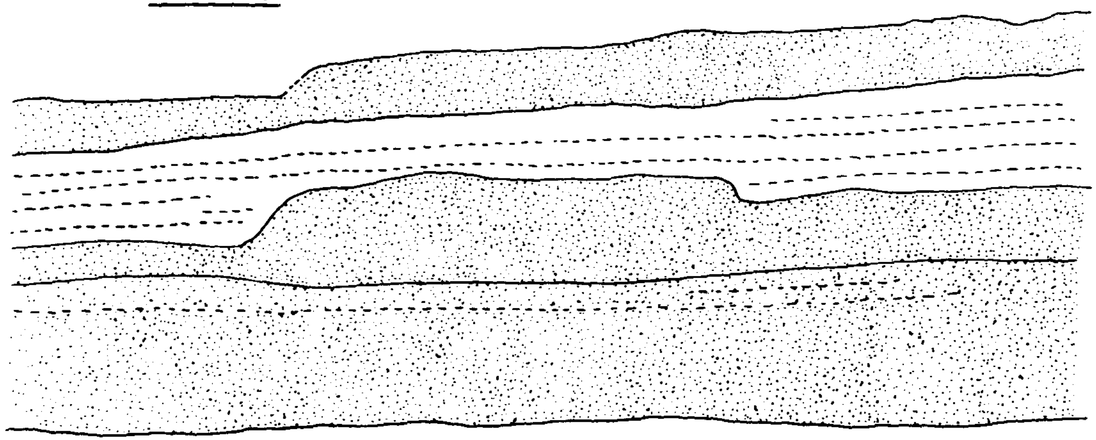


FIG 5.16

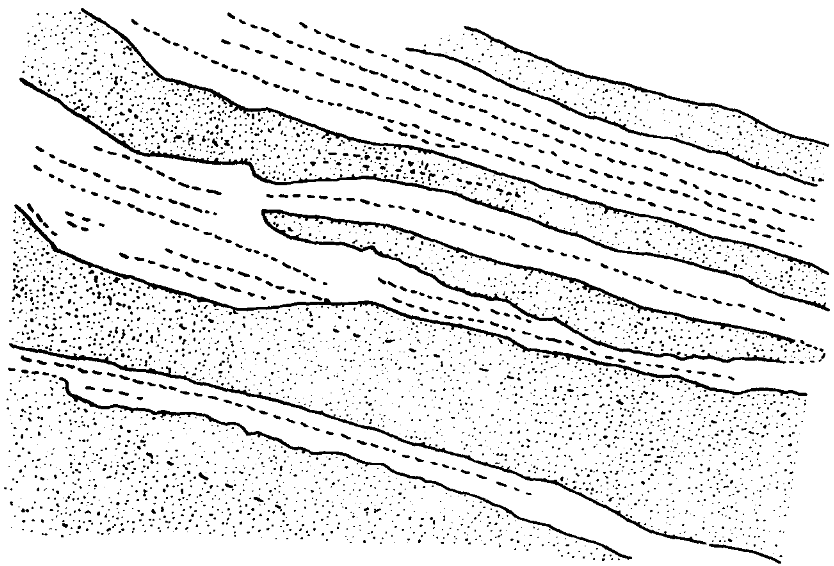
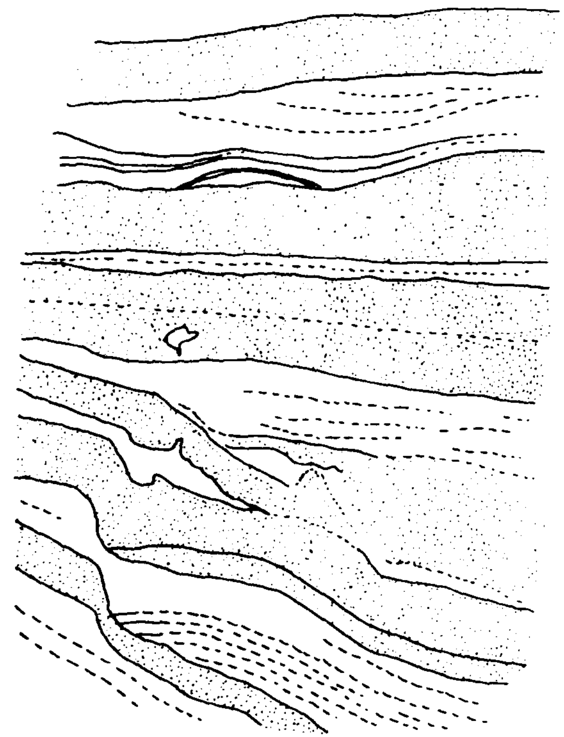


FIG 5.17



combined with sand liquefaction may account for the irregular tops of many of the beds.

Therefore this facies is dominated by T_a based turbidites, including T_{ad}, T_{abd} and T_{abcd} sequences. Very coarse grained beds are rare and amalgamation is generally uncommon. Liquefaction/loading structures suggest this facies was rapidly deposited.

Facies 2.

This facies is dominated by siltstones and mudstones which are usually parallel laminated. The siltstones may be interbedded with relatively thin bedded, usually T_b or T_c based sandstones (of the order of 10–30cm thick), though the sandstone-shale ratio is generally low (c.0.3 and always less than 1). The siltstones and mudstones show a range of colours: green is the most common, grey is more common in the upper and lower parts of the section, purple, red and yellow. Green and red shales are often interbedded or interlaminated in some parts of the section whereas elsewhere shales of a particular colour predominate. In general the contacts between shales of different colours are laterally continuous, though in detail there may be lateral variation. There is an increased proportion of ferric iron in red shales relative to green shales, which are richer in ferrous iron. Purple shales have an intermediate ferrous-ferric iron content (Hagemann 1957). Thus the main control on the colour of the shales is the oxidation state of iron within these mudrocks.

The occurrence of red shales within marine sediments may be accounted for in two main ways:

a) The red colour, due to haematite staining was inherited from the source area at a time before the evolution of land plants when oxidised iron was concentrated on most land surfaces. During periods of transgression e.g the early Cambrian this ferric-stained sediment might be transported to depositional basins. If the amounts of organic matter within the sediment were low and sedimentation rates were high (conditions which were probably met in Cilan Grit times) then reduction of the sediment might have been prevented and the sediments would have kept their red colour (Ziegler & McKerrow 1975). This might imply that the green and the red shales had different sources. However, the lateral discontinuity of bands of a particular colour suggest that the colour is at least in part diagenetically controlled. The presence of green shales may therefore represent shales which had a slightly higher organic content, the decay of which allowed reduction in the shales from ferric to ferrous iron.

b) Turner (1978) interpreted red shales from the Caerfai Series of South Wales as acquiring their red colour from authigenic haematite, formed from intrastratal solutions of iron-bearing minerals. This may be an important process in the formation of red shales within the Cilan Grits.

The mudstone-siltstone facies also contains intraclasts, some of which have picked up sand grains on their outer surfaces. Locally the facies contains irregular lenses, ?sedimentary intrusions and chaotic downloaded pillows of sandstone.

Trace Fossils.

On the foreshore west of Trwyn Cilan large fallen blocks of Cilan Grit contain four types of trace fossil:

i) Circular sandstone-filled depressions up to 8cm wide, averaging 3-4cm and a few millimetres deep. They appear to be the resting trace of a soft bodied animal (*Bergaueria*).

ii) Horizontal burrows 1-3mm wide and 1-5cm long (*Planolites*). Sometimes the high density of horizontal burrows results in a mat-like appearance on certain bedding planes.

iii) Branching horizontal burrows (c.f. *Phycodes*).

iv) Vertical burrows of a similar width to the horizontal burrows (*Skolithos*). Many burrows appear to be paired (*Arenicolites*).

Crimes (1970b) also records *Sinusites*, intertwining burrows and burrow networks, the latter representing grazing or "farming" traces (c.f. *Palaeodictyon*). Trace fossils are generally most common in the siltstone-mudstone facies, probably as a result of more abundant nutrient supply and increased preservation potential.

Conclusions.

The Cilan Grits show different types of sequence, ranging from sand-rich in facies A and B to more mud-rich in facies 2. The Cilan Grits contain features indicative of Mutti's (1979) B1,C1 facies with some beds more typical of B2 and C2 facies. The lack of obvious channelisation apart from single cut and fill events, as well as an absence of consistent fining and thinning upward sequences indicates non-channelised turbidity current flow. Cross-bedding indicates traction current activity. The predominance of T_a based Bouma sequences and generally high sandstone-shale ratios are indicative of relatively proximal turbidites. The occurrence of pebble beds, multiple and reverse grading and scours seem to indicate that facies B was deposited from high density turbidity currents, whereas more complete Bouma sequences in facies 1 and 2 indicate more dilute flow.

The Cilan Grits differ in several respects from the Barmouth Formation, their lateral equivalents in the Harlech Dome:

1) Although both contain similar facies, the siltstone-mudstone facies with thin bedded sandstones is well developed in the Cilan Grits especially near the top and is uncommon in the Barmouth Formation.

2) In the Barmouth Formation the Amalgamated Coarse Grained Facies (equivalent to facies B of the Cilan Grits) is generally much more common.

3) The Cilan Grits were derived from the northeast in contrast to the Barmouth Formation which was derived from the south.

4) The Cilan Grits overall are thicker (300m) compared to the Barmouth Formation (less than 200m) which suggests that the upper and/or the lower boundaries of the respective

units are probably not synchronous.

These features and particularly the large difference in palaeocurrents indicate that the Cilan Grits was deposited in a different turbidite system to that of the Barmouth Formation.

5.5 : Caered Mudstones and Flags.

The Caered Mudstones and Flags are over 150m thick and are exposed near Trwyn Llech-y-doll and on the western side of Porth Ceiriad. Most of the outcrops are fault-bounded so only minimum thicknesses can be given for the unit as a whole or for subdivisions within this unit. The Caered Mudstones and Flags were subdivided by Nicholas (1915) and are described below:

1) Lower Caered Mudstones (over 34m).

West of Trwyn Llech-y-doll [SH 2995 2336] they are comprised of thick dark green mudstones interbedded with thin bedded (generally less than 10cm thick) parallel laminated grey and green siltstones. Occasional 5-10cm thick siltstones contain ripple cross lamination. Thus the sequence is dominated by thin bedded T_b,T_c and T_d based dilute turbidites. Simple horizontal trace fossils also occur.

Rare thicker bedded sandy turbidites occur near the base indicating that occasional proximal T_a based turbidites were still being deposited. Almost directly above the Cilan Grits there is a 70cm thick, hard, green, fine grained, siliceous bed which has a subconchoidal fracture. This bed is overlain by a 10-15cm thick sandstone bed, the lower boundary of which is highly irregular and contains isolated lenses of fine grained siliceous sediment. Nicholas (1915) interpreted the flinty bed as a tuff, and the presence of angular quartz grains with crystal faces and re-entrant angles supports this. The bed is poorly laminated but probably represents the deposition of relatively distal airfall of homogenous volcanic dust through the water column. Current activity was probably negligible at this time. At some stage relatively soon after the deposition of the tuff a turbidite sandstone was deposited on top, which resulted in the tuff turning thixotropic. As a result the sand down-loaded into the underlying tuff. This was not an isolated volcanic event since thin (up to 14cm thick) siliceous tuff beds occur interbedded with the mudstones above, though they decrease in abundance and thickness upwards.

2) <u>Caered Flags</u> (over 35m).

The base of the Caered Flags is not exposed so the true thickness is not known. The Caered Flags in Porth Ceiriad contain red and green fine grained sandstones which are up to 100cm thick, though most are between 5-20cm thick. Parallel laminated siltstones however predominate and mudstones are also common especially in the upper part of the section where they pass up into the Upper Caered Mudstones. Some thin bedded sandstones are present, a few of which contain ripple cross lamination as part of T_{cd} sequences. The ripples have an amplitude of 1-4cm and a wavelength of 7-15cm. Trough cross lamination is most common and both starved and climbing ripples occur. Palaeocurrents obtained from cross lamination are variable, though flow from the southeast dominates (Fig 5.18). The generally low sandstone-shale ratios and dominance of base absent Bouma sequences indicates deposition from mainly dilute turbidity currents.

3) <u>Upper Caered Mudstones</u> (over 84m).

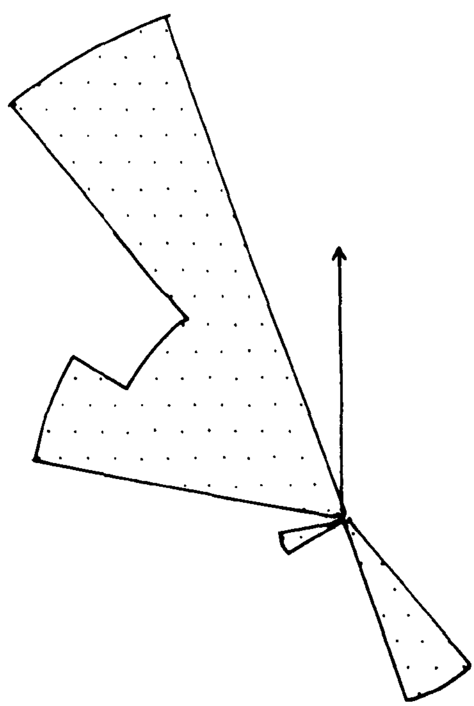
This unit is dominated by blue and green mudstones, some of which are slightly manganiferous. The mudstones also contain thin, light-coloured calcareous laminae and thin beds. Agnostid trilobites, *Paradoxides* (indicative of a Middle Cambrian age) and inarticulate brachiopods are also present. The top of the unit is not exposed but is presumed to be overlain by the Nant-pig Mudstones, a sequence of dark, organic-rich mudstones.

<u>Conclusions.</u>

The lower part of the Caered Mudstones and Flags is mainly composed of base absent Bouma sequences which represents a period of dilute turbidity current deposition. The Upper Caered Mudstones, however may have been deposited in a relatively quiet water, shelf environment, since there

FIG 5.18 Caered Mudstones and Flags: Palaeocurrents.

cross lamination:



n=17

is an increase in the abundance of faunal activity. These mudstones and the overlying Nant-pig Mudstones are similar to the Menevian shelf deposits of South Wales (Nicholas 1915). Some of the thin graded units may have been storm generated. Therefore the sequence as a whole may shallow upwards. The upper contact of the Nant-pig Mudstones is disconformable and overlain by stromatolites (which provide strong evidence for shallow water) and Maentwrog shelf storm deposits (Bose 1983) so a regressive trend in the Upper Caered Mudstones is probable.

5.6 : Conclusions.

The Lower and Middle Cambrian of St Tudwal's peninsula contains both thick bedded, sand-rich turbidites and thin bedded silty turbidites, though over most of the succession the former predominates. It is difficult to detect any general trend in these deposits; instead pulses of coarser or finer sediment, probably controlled at source seem to be the major influence on vertical trends. There is however evidence for shallowing near the top of the sequence. Laterally beds appear to be relatively continuous along strike, though good lateral control is possible only in the Hell's Mouth Grits. This suggests a non-channelised geometry for most beds, though individual scour and fill events were not uncommon in parts of the sequence. Turbidites below the top of the Cilan Grits are primarily derived from the northeast, while units above this contact (including the Upper Cambrian) are sourced from the southeast and south. So there was a profound change in depositional setting above the top of the Cilan Grits.

The Harlech Grits Group of the Harlech Dome and the sequence on St Tudwal's from the Hell's Mouth Grits to the Upper Caered Mudstones are overall very similar. The upper Middle and Upper Cambrian rocks of the two areas are also similar, particularly the Nant-pig Mudstones and the Clogau Formation of the Mawddach Group. The close lithological correlation between the two areas suggests that both St Tudwal's and the Harlech Dome were deposited within the same basin. Although the coarse grained turbidites of the Hell's Mouth Grits and Rhinog Formation were deposited at the same time and the Cilan Grits and Barmouth Formation were synchronous, sufficient differences occur in facies and palaeocurrents for them to be assigned to different turbidite systems. However the fact that sand-rich turbidites were deposited at similar times in the two areas indicates a common external control on the supply of sediment to the basin.

CHAPTER SIX

ARFON

CHAPTER SIX : ARFON.

6.1 : Introduction.

Arfon is the area of mainland North Wales to the southeast of the Menai Straits and Anglesey. Strata of mainly Cambrian age are exposed in the predominantly lowlying ground northwest of the Snowdonian massif. Cambrian rocks occur in two inliers: the Bangor Ridge (between Bangor and Caernarvon) and the Padarn Ridge (Bethesda-Llanberis-Nantlle), (Fig 6.1). The Lower and Middle Cambrian sequence is comprised of ash-flow tuffs at the base followed by a mixed sedimentary and volcanic succession which is overlain by a thick sequence of slates and then by sand-rich turbidite sandstones.

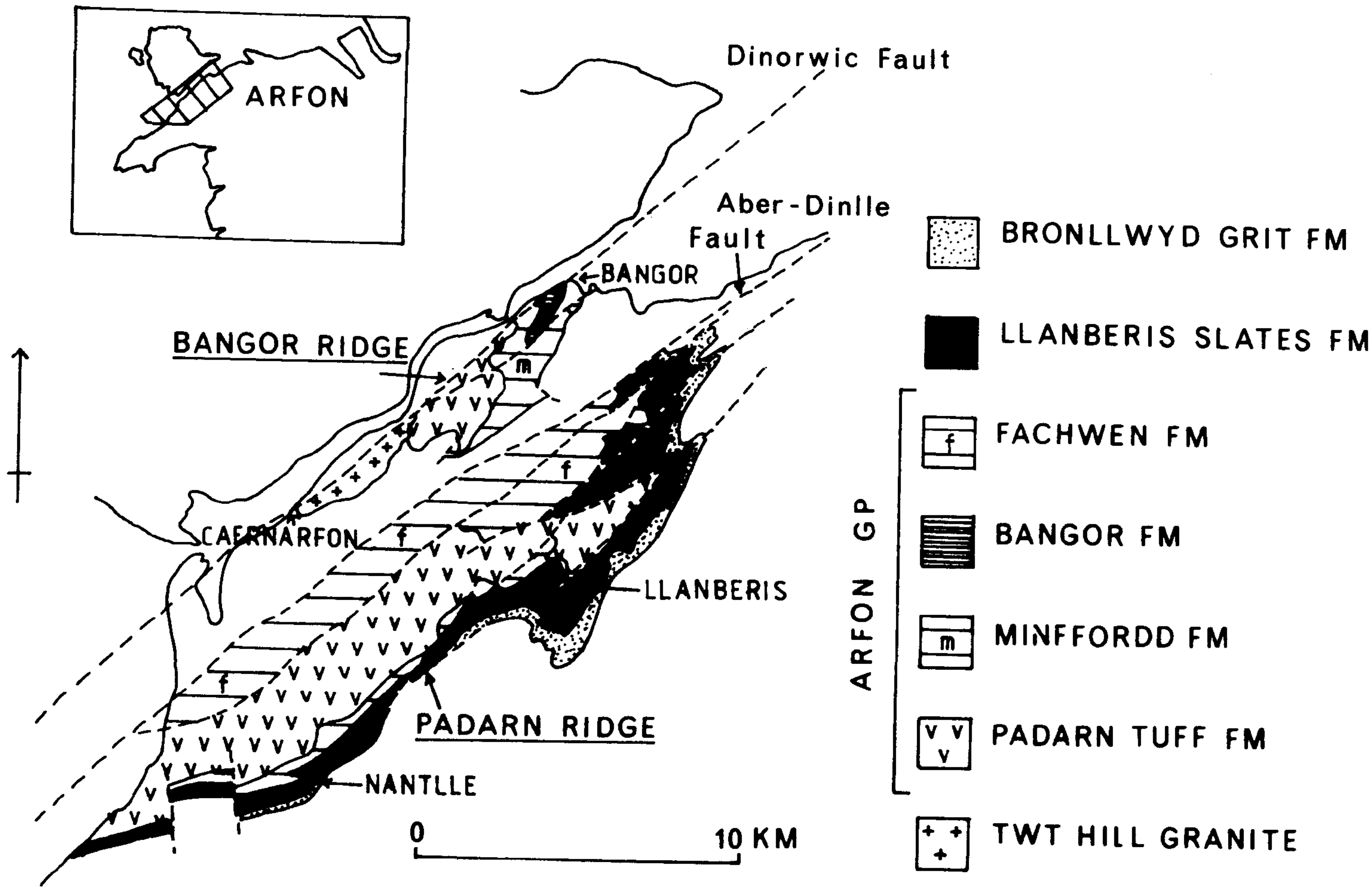
Defining the base of the Cambrian sequence in Arfon has been a matter of controversy. The oldest strata of Arfon are included in the Padarn Tuff Formation which was previously regarded as intrusive (Ramsay 1866). However Bonney (1879) suggested that it was extrusive and petrographic (Rast 1962; Dearnley 1967) and field studies indicated that it was an ignimbrite, deposited from ponded pyroclastic flows (Reedman *et al.* 1984). The nature of the contact between the tuff and the overlying conglomerates in the Llanberis area has been hotly debated, with polarisation into two main views:

1) The contact is conformable, based on Ramsay's observation that there appears to be a transition from one lithology to another.

2) The contact is unconformable and represents the base of the Cambrian in the Llanberis area. Unconformities were identified at, what were shown by Wood (1969) to be, separate conglomeratic horizons (Blake 1888, 1893, 1898; Bonney & Raison 1894; Geikie 1891; Green 1885; Hicks 1878).

The confusion was compounded by the rapid facies changes within the lower part of the succession in the Llanberis area (Wood 1969), the variable nature of the tuff-conglomerate contact from conformable to unconformable

FIG 6.1 Geological Map of the Lower and Middle Cambrian rocks of Arfon.



(after Reedman et al. 1984)

and the occurrence of several tuff horizons above the Padarn Tuff. Wood used the above evidence to suggest that there was little or no appreciable time gap between the deposition of the two units. Trilobites were found at the top of the slates (Woodward 1888; Howell & Stubblefield 1950) which indicate a Lower Cambrian age. The sequence below this is unfossiliferous so the Padarn Tuff Formation which occurs 1500m below the fossil horizon may be Lower Cambrian or late Precambrian.

The lower part of the Cambrian sequence is very variable laterally, even within a relatively small area (see Wood 1969). Substantial differences occur even within the Padarn Ridge, for instance between Llanberis (Wood 1969) and Nantlle (Hughes 1917; Morris & Fearnsides 1926; Cattermole & Jones 1970). However the upper part of the Cambrian succession in these areas and the area north-east of Llanberis (Williams 1923, 1930; Evans 1968) is more laterally continuous. Lateral changes also occur in the Lower Cambrian between the Padarn and Bangor Ridges (Blake 1892; Greenly 1944, 1945). The succession in each area have been divided into different stratigraphic divisions (BGS 1:50,000, Sheet 106; Howells *et al.* 1985). The lateral differences can be accounted for by syn-sedimentary fault movement on the Aber-Dinlle Fault which outcrops between the Bangor and Padarn Ridges. Smaller scale syn-sedimentary faulting was also active during deposition of the Llanberis Slates Formation and Bronllwyd Grit Formation (Webb 1983).

The two inliers of the Arfon area, the Bangor and Padarn Ridges will be described separately.

6.2 : Bangor Ridge.

The Bangor Ridge is the most northerly of the two main Cambrian inliers of Arfon. The Bangor inlier is aligned NE-SW; it is 12km long with a maximum width of 3km (Fig 6.2). The following succession is exposed in this area (Fig 6.3; Reedman *et al.* 1984; Howells *et al.* 1985):

1) Padarn Tuff Formation. Greater than 800m of acid ash-flow tuffs.

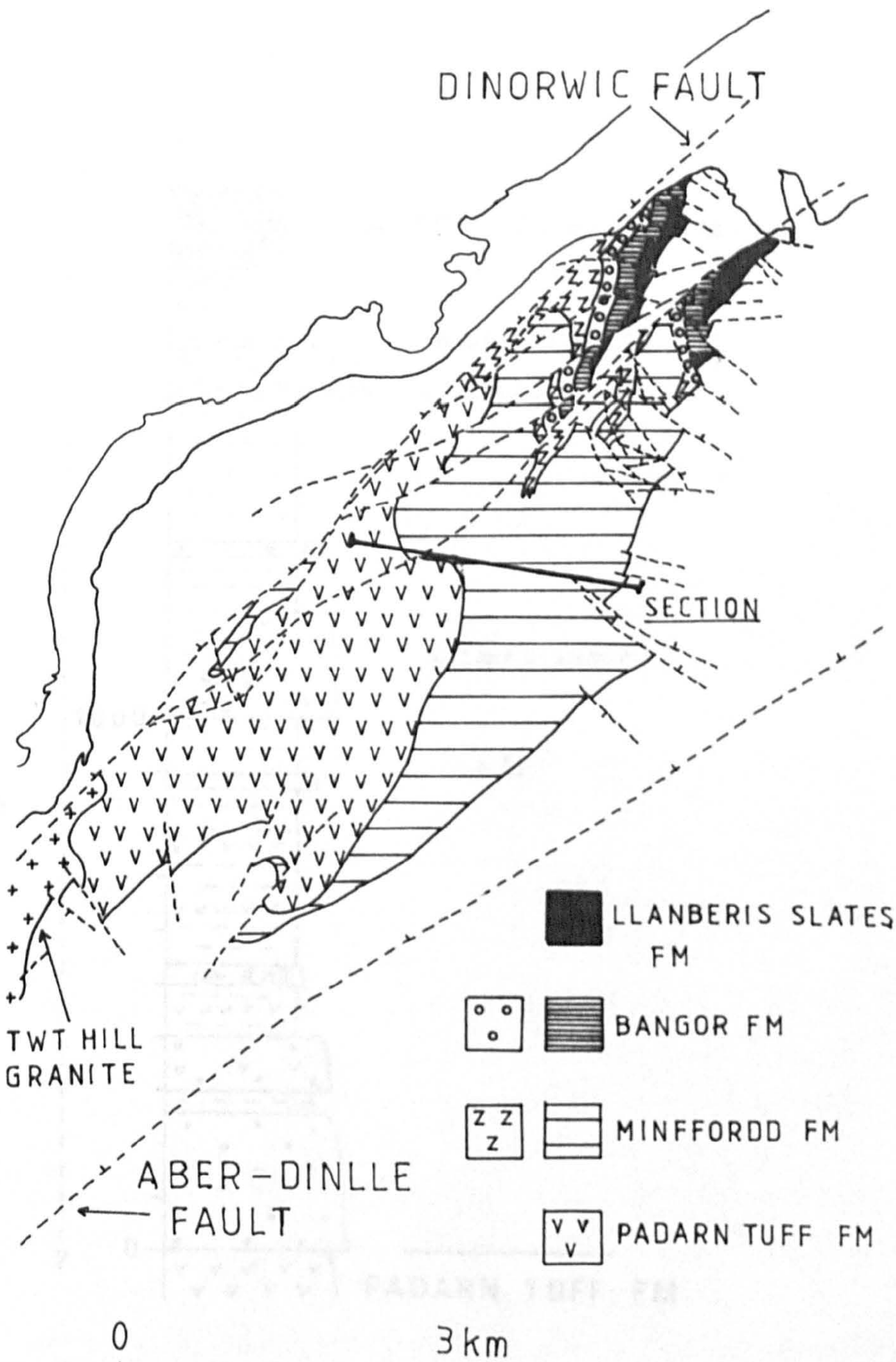
2) Minffordd Formation. Over 1500m of predominantly siltstones, acidic air-fall tuffs and tuffites interbedded and interlaminated with volcanic-derived conglomerates, sandstones and thin ash-flow tuffs. A thicker acid tuff occurs in the upper part of the formation (BGS 1:50,000 Sheet 106).

3) Bangor Formation. Approximately 240m thick, lying unconformably on tilted Minffordd Formation. The Bangor Formation is comprised of conglomerates in the lower part (Siliwen Conglomerate member) and interlaminated sandy and silty acid tuffs and tuffite (Mountain Tuffite Member) in the upper part.

4) Llanberis Slates Formation. Approximately 100m of red and purple siltstones are exposed. The full thickness is not exposed due to pre-Arenig erosion and overstep by the Arenig Graianog Sandstone. The Arenig sediments in this area rest unconformably on Cambrian and overstep strata ranging from the Padarn Tuff Formation in the southwest to the Llanberis Slates in the northeast.

A west to east transect across the Bangor Ridge from Penrhos-garnedd to Caerhun was studied, using sections exposed in cuttings on the newly constructed Bangor bypass (Fig 6.4). In this part of the bypass strata ranging from the Padarn Tuff Formation to the upper Minffordd Formation are exposed. The exposure here is good but not continuous and it is difficult to unravel the exact succession because of faulting. A major NE-SW trending fault was recognised by

FIG 6.2 Geological Map of the Bangor Ridge.



(after Reedman et al. 1984)

FIG 6.3 Geological Succession, Bangor Ridge.

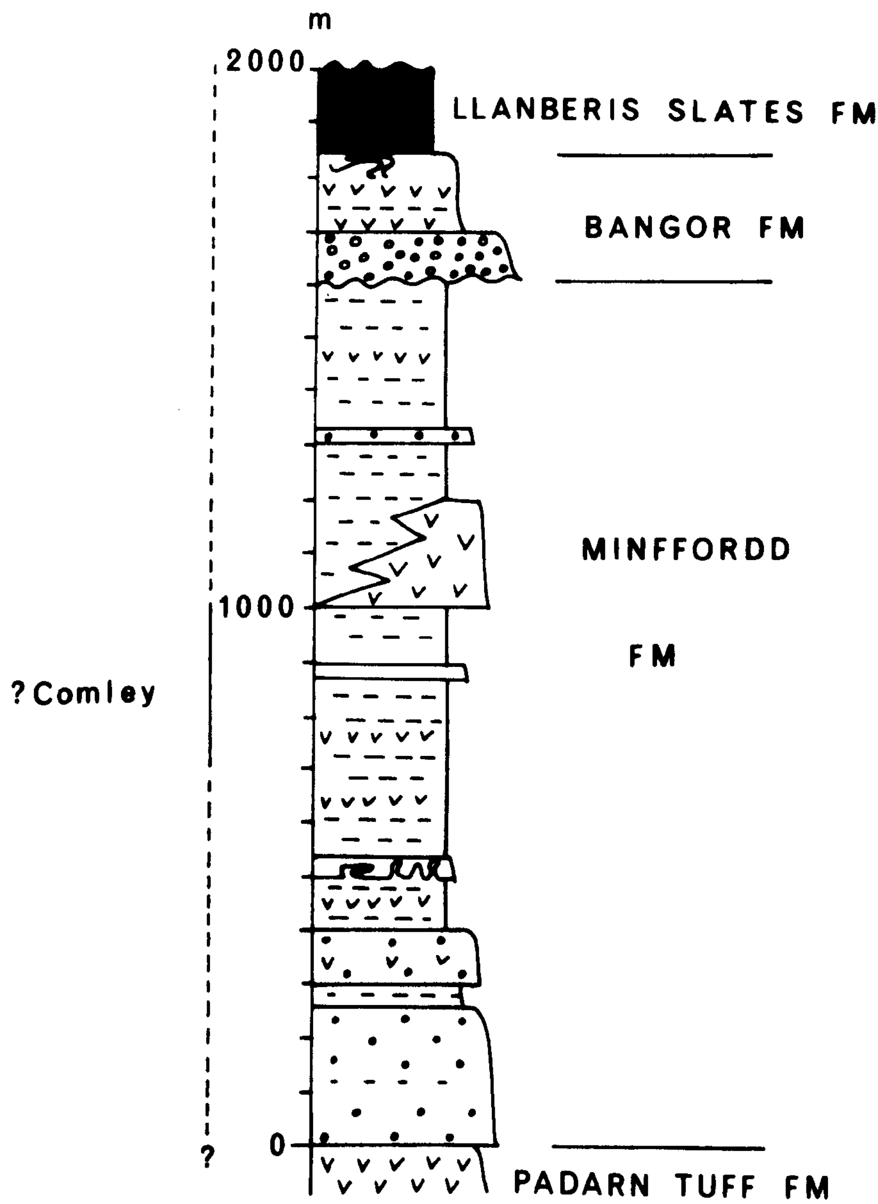
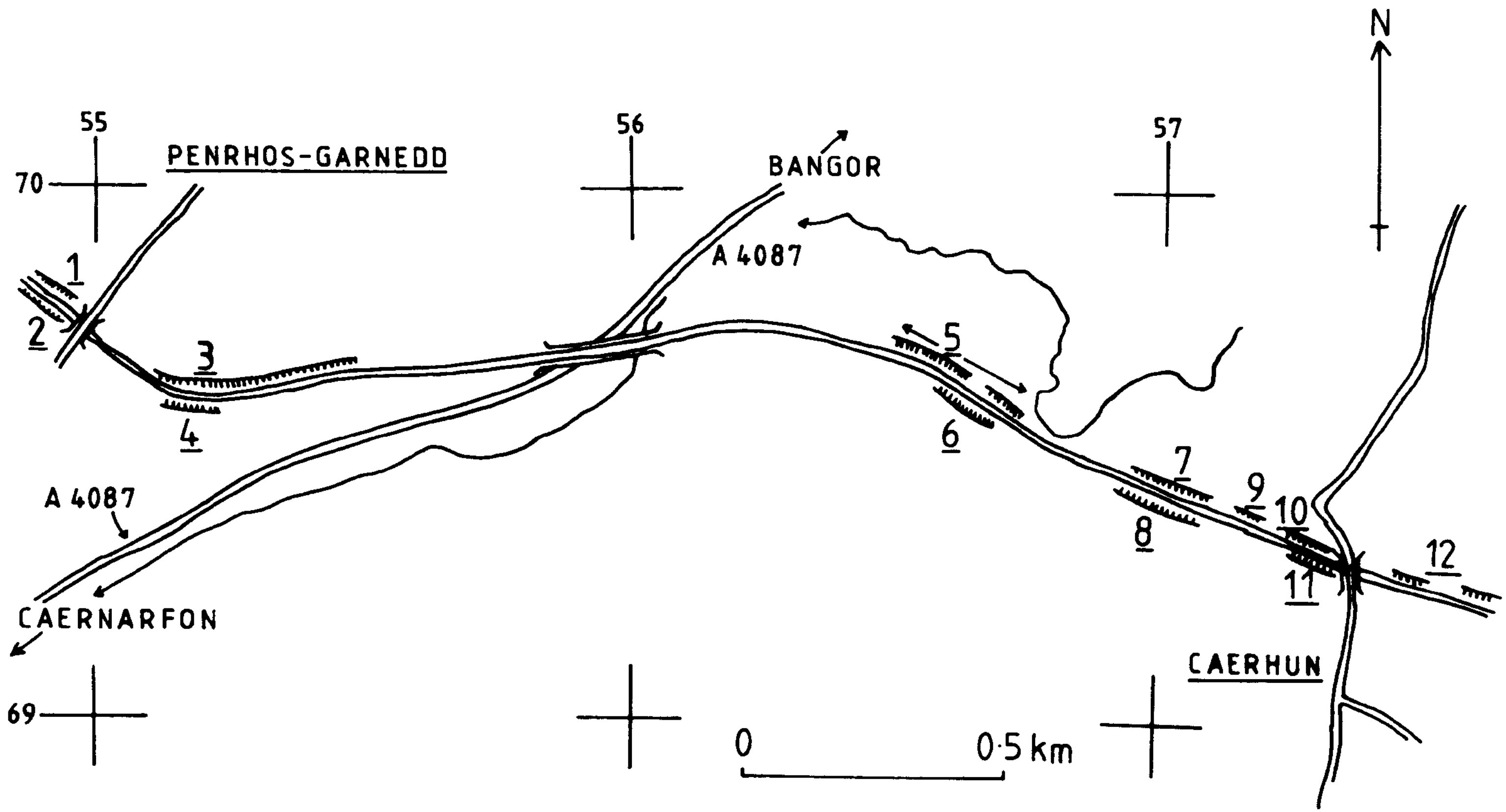


FIG 6.4 Locality Map, Bangor Bypass.



the Survey (BGS 1:50,000 Sheet 106) which has a downthrow to the northwest and occurs between sections 3 and 4 and sections 5 and 6.

1) Padarn Tuff Formation.

The Padarn Tuff Formation outcrops in the western part of the bypass, in sections 1, 2 and 3. It is predominantly massive, with little indication of bedding, though it is well jointed. The tuff is porphyritic with phenocrysts up to 4mm across (mean 1mm) of lamellar and simple twinned feldspar and euhedral to subhedral quartz, contained within a very fine grained crimson-coloured groundmass. Many of the phenocrysts have highly embayed, angular outlines and some show re-entrant angles as a result of twinning. There are also irregularly shaped clasts which are probably shards, indicating that this lithology has a pyroclastic origin. Pumice fragments and evidence for welding are generally lacking. There are few vertical changes, so identification of separate flow units is not possible. The massive nature of the deposit indicates that the Padarn Tuff Formation was deposited from ponded ash-flows (Reedman *et al.* 1984).

2) Minffordd Formation.

Several facies occur within the Minffordd Formation:

A) Coarse, red volcanic sandstones.

B) Coarse, green volcanic sandstones.

C) Volcanic sandstones interbedded with fine grained tuffs.

D) Fine grained tuffs.

E) Siltstones.

A) Coarse, red volcanic sandstones. This facies is comprised of predominantly massive, thick bedded coarse to very coarse sandstones. The sandstones directly overlie the Padarn Tuff Formation and outcrop in the eastern part of section 3. The sandstones contain grains up to medium pebble size and are commonly made up of alternating c.1cm thick bands of very coarse and coarse to medium sandstone. This lamination is usually diffuse, though any particular lamina is usually well sorted. There are also occasional heavy mineral layers (of haematite/magnetite).

This facies is lithic-rich, the most common clasts being of red-brown, siliceous ash-flow tuffs similar to the underlying Padarn Tuff Formation, thus indicating that this facies was mainly derived locally. Other clast types are mainly volcanic-derived: spherulitic quartz, welded ash-flow tuffs and angular phenocryst-derived quartz and feldspars, while other clasts appear to be of intraformational origin. Some clasts, however, were derived from more distant sources: micrographic fragments (indicating a granitic source) and probable metamorphic quartz containing grains with highly undulose extinction and sutured grain contacts. This facies contains a low proportion of matrix- the grains are often cemented by thin chlorite rims and many of the volcanic or volcanic-derived lithic fragments have deformed on compaction to fill existing pore spaces. Interbedded with the sandstones are rare, thin beds of red, cherty mudstones which contain occasional scattered sand grains. In general the sandstones get finer up the sequence.

Since this facies overlies subaerial ash-flow tuffs and the sediment is well sorted but relatively immature mineralogically and in general fines upwards it was probably deposited in a fluvial setting, possibly within a coarse braided channel system.

B) Coarse green volcanic sandstones. This facies is dominated by massive, thick bedded volcaniclastic sandstones. The sandstones outcrop in the western part of section 5 where 10m are exposed. These sandstones are rich in volcanic lithic fragments including both welded and non-welded tuffs, but these appear to be more basic and quartz-poor relative to facies A, which occurs stratigraphically beneath facies B. Grain sizes vary from granule conglomerate to medium sandstone, though coarse sandstone is most common. In general the coarser beds are richer in lithics, whereas the finer beds are more feldspathic. The matrix is fine grained, grey in colour and contains some euhedral quartz crystals which were probably originally phenocrysts. The occurrence of clasts of ashflow tuffs (some with shards), chert and basic volcanics showing a trachytic texture, indicates, at least in part, an epiclastic source for these sandstones, and therefore this facies appears to be dominated by coarse grained tuffites. These tuffites are interbedded near the top of the sequence with parallel laminated fine grained tuffs.

The abundance of relatively immature volcanic fragments indicates deposition near to source and there may be some primary volcanic material within this deposit. The occurrence of thick massive beds containing abundant matrix indicates deposition from relatively high density flows (?turbidity currents).

C) Volcanic sandstones interbedded with fine grained tuffs. This facies overlies facies B in section 5. Lithologically the sandstones are similar to those in the underlying facies except that they lack the primary volcanic matrix. Some beds are graded; one example is described in detail. It contains clasts of grey silt, black and grey volcanics, feldspar,quartz, and chert. Overall this bed grades up from coarse to fine sandstone. However the basal 4mm of the bed shows inverse grading from a relatively matrix-rich coarse to medium sandstone to a matrix-poor coarse sandstone above. The bed shows three laminae of inverse graded sandstone, approximately 1.5cm thick, though some are more diffuse and more laterally variable than others. In the top few centimetres the bed grades rapidly up into convolute laminated fine sandstone. The presence of small scale inverse grading and thin finer laminae, interpreted as shear laminae are indicative of high density flows (Lowe 1982).

Parallel laminated sandstones are occasionally present as is cross lamination (1-2cm amplitude), though convolute lamination is slightly more common. Intraclasts are common locally (and may show small scale sand injection) and where concentrated may form intraclast breccias. Massive (T_a) beds are abundant and commonly have highly loaded bases with up to 60cm amplitude, often with associated flame structures. Soft sediment deformation features are particularly common and include chaotic load and injection structures. These structures include ball and pillow structures as well as "intraclasts" which appear to have been derived from injection of sand around the finer grained sediment. Loading has produced beds with highly irregular geometries which have undulose and intrusive bases and tops. Slump folding is also present locally. Microfaulting is also common, which is probably syn-sedimentary in origin.

Some beds contain wispy laminations composed of blue-black squashed pumice (fiamme). They may occur scattered throughout volcanic siltstone beds or may form the main proportion of the sediment in beds up to 8cm thick. The fiamme are usually about 0.5mm thick and 1mm long (maximum 8mm). They commonly have cuspate outlines, which may have formed due to vesiculation. The presence of shards in the beds between the fiamme-rich bands indicates a probable ash-fall origin for these beds, though it is possible that some pumices were welded in thin ash-flows.

The volcanic sandstones are interbedded with thin bedded fine grained tuffs which are of similar type to the overlying facies D.

The abundance of shards and fiamme in many relatively thin beds indicate that they were probably deposited from ash-fall eruptions. These beds are intercalated with thicker bedded volcanic sandstones which are also largely composed of volcanic debris but contain multiple and inverse grading, structures typical of high density turbidity currents.

D) Fine grained tuffs. This is the most common facies in the Minffordd Formation; it is particularly common in the upper part of the sequence and occurs in sections 5-11. Most beds are flinty with a conchoidal to subconchoidal fracture and are characteristically parallel laminated with few other structures apart from rare sets of low amplitude cross lamination and microfaulting. Microfaulting is locally very common and results in displacements ranging from 1mm to a few centimetres. The faults may be normal or reverse and the distribution of fault types is usually irregular, though they often show a braided pattern in vertical sections. However some examples from loose blocks were found which show tensional (normal) faults perpendicular to compressional (reverse) faults, though their *in situ* orientation is not known. Faulting probably occurred relatively soon after deposition since in some cases there appear to be thickness changes across faults and occasionally there is spillage of sand along the fault plane. The latter may indicate that at least some of the sediment was poorly consolidated, though much of the faulting seems to indicate a relatively brittle response to deformation. In one example a set of laminae were tilted as fault blocks about 1cm across and overlain by undisturbed laminae, which seems to represent a micro-unconformity. Thus a syn-sedimentary origin for the microfaulting is most likely.

Several types of beds occur in this facies:

i) Thinly laminated fine grained tuffs- the dominant bed type, which characterises this facies. Sand laminae are uncommon; instead laminae are composed of alternating lighter and darker green massive tuffaceous laminae. Some contain chloritic fragments or one-grain thick laminae of feldspar crystals, which may have been deposited from distal ash-falls. One example was found of an isolated large chloritic fragment (11mm diameter), which downloaded into the laminae beneath and was draped by the laminae above. This feature is probably a dropstone-like structure produced by a volcanic bomb.

ii) Thinly laminated fine grained tuffs, but with sand laminae. One example, taken from the lower part of the facies in section 5 contains laminae (1-2mm thick) of dark mudstone, greenish to light grey siltstones/mudstones, dark grey siltstones/mudstones and light coloured ashy fine sandstone laminae.

iii) Thin lensoid sandstone beds, usually of fine or medium sandstone which are usually of the order of 1-2cm thick. They may be graded, but frequently have abrupt tops. They are very variable in thickness, are often loaded and sometimes have small scale sand intrusions associated with them. Thin lenses of intraclast breccia also occur.

iv) Thicker bedded sandstones. These beds may be graded or ungraded. They are usually massive and are commonly about 30cm thick, though they have a large range. Several types occur, including fine grained, green, chloritic sandstones and medium grained greywacke-type sandstones (volcanic-poor). Sandstones may occur randomly within the sequence or may show a slight clustering, though even here sandstone-shale ratios rarely exceeds 1.

v) Very thick massive beds, transitional with bed type "iv" and is interbedded randomly within the sequence. Type "v" beds are usually fine to medium grained. However in section 10 a 159cm thick bed of granule conglomerate to very coarse sandstone outcrops. This bed contains clasts of green, grey and red siltstones/mudstones and overall is normally graded. This sandstone contains angular volcanic-derived clasts as well as basic volcanic clasts showing a trachytic texture; this bed is therefore a coarse tuffite.

Thin beds/laminae containing abundant angular feldspar grains indicate deposition from ash-fall eruptions. Graded beds/laminae of alternating sand and fine grained tuff indicate deposition from dilute turbidity currents. Soft-sediment deformation indicates rapid rates of sedimentation and may have been induced by seismic activity. The proportion of volcaniclastic sediment decreases upwards and thick massive beds higher up in the sequence tend to be more mature mineralogically and were probably deposited from higher density, sand-rich turbidity currents.

E) <u>Siltstones.</u> This facies is composed of parallel laminated siltstones. They occur in section 12, in the upper part of the sequence and overlie strata belonging to facies D. The siltstones pass up from green slates into grey, purple and red slates and are similar to and possibly transitional with the Llanberis Slates Formation. This facies is overlain disconformably by Arenig sandstones in section 12.

These siltstones may have been deposited in a quiet water basinal setting by analogy with similar deposits in the Padarn Inlier.

3) Conclusions.

Deposition within the Arfon Basin may have been volcanogenically controlled (Reedman *et al.* 1984). The sequence exposed along the Bangor bypass represents a passage upwards from subaerial ash-flow tuffs (Padarn Tuff Formation) to possible alluvial fan/braided fluvial deposits of the early Minffordd Formation (facies A). The lower Minffordd Formation is mainly derived from erosion of acidic Padarn Tuff-type source rocks. However thick, massive beds of facies B, which represent coarse, high density flows and overlie facies A sediments were derived from a more basic source. Reedman *et al.* (1984) suggests that these sandstones may have been derived from basic volcanics in the Gwna Group of Anglesey. Facies B deposits also contain evidence for contemporaneous volcanic activity and fine upwards into interbedded sandstones and fine grained tuffs (facies C). Facies C was deposited from turbidity currents and shows soft-sediment deformation and microfaulting. The former may have been produced by liquefaction of sand beds, which may have been triggered by fault movements at the basin margin. Reedman *et al.* found sponge spicules in the upper part of the Minffordd Formation which suggest that at least part of the formation was deposited in marine conditions. Facies fines upwards into distal tuffs (both air-fall and redeposited fine-grained volcanic material) of facies D. The primary volcanic input also decreases upwards through the sequence; graded quartz-rich greywacke sandstones indicate non-volcanic deposition in a relatively deep basin. These sandstones pass upwards into facies E which may be transitional with the Llanberis Slates Formation.

In the northern part of the Bangor Ridge there is evidence for overlap of the upper part of the Minffordd Formation onto the Padarn Tuff Formation (Reedman *et al.* 1984) indicating possible fault movement on the Dinorwic Fault. There is therefore a slight unconformity between the Minffordd Formation and the overlying Bangor Formation. The

Siliwen Conglomerate member of the Bangor Formation is composed mainly of Padarn Tuff-type clasts as well as occasional clasts of Monian schists (Greenly 1923) which may indicate penecontemporaneous uplift and erosion of the Mona Complex and Padarn Tuff on Anglesey. Deposition occurred within small fans in a narrow fault-confined Arfon Basin (Reedman *et al.* 1984). The overlying Mountain Tuffite Member is comprised of distal turbidites and tuffites. Slump folding and sediment disruption within this member may also have been induced by fault movements within the basin. The Bangor Formation is overlain by red siltstones of the Llanberis Slates Formation, which can be correlated with the Llanberis area, in contrast to the considerable lateral variations that occur below this formation. The continuity of sequences above the Arfon Group may indicate a change from localised fault controlled sedimentation during deposition of the Minfordd and Bangor Formations to more uniform conditions while the Llanberis Slates Formation was being deposited. However only the lower part of the Llanberis Slates Formation is exposed in the Bangor area and it is not clear whether the slates were of similar thickness to Llanberis and were removed by pre-Arenig erosion or whether the formation was thinner in the Bangor inlier. Whether the overlying Cambrian formations at Llanberis had lateral equivalents at Bangor is not known. A combination of thinning and pre-Arenig erosion seems the most likely explanation, though at what time or times the Cambrian succession of the Bangor inlier was eroded is difficult to determine. However it seems likely that sedimentation and probably later erosion of Arfon Basin sediments in the Bangor inlier was controlled by movement on the Dinorwic Fault to the northwest and the Aber-Dinlle Fault to the southeast.

6.3 : Padarn Ridge.

The Padarn Ridge is the more southerly of the Cambrian inliers of Arfon. The Lower and Middle Cambrian part of the inlier is 25km long and up to 9km wide and is aligned NE-SW between Nantlle, Llanberis and Bethesda (Fig 6.1). The following succession is exposed in the Llanberis area of the Padarn Ridge (Fig 6.5) (Reedman *et al.* 1984; Howells *et al.* 1985):

ARFON GROUP. (Formations 1 and 2)

1) Padarn Tuff Formation. A sequence of massive, acidic ashflow tuffs.

2) Fachwen Formation. A mixed sedimentary sequence of conglomerates (especially common near the base), sandstones, siltstones and thin beds of tuff and tuffite.

3) Llanberis Slates Formation. Predominantly cleaved siltstones and silty mudstones with occasional sand packets.

4) Bronllwyd Grit Formation. Mainly sand-rich packets of turbidite sandstone.

The succession at Nantlle (Fig 6.6) (Morris & Fearnsides 1926; Cowie *et al.* 1972) differs considerably from that at Llanberis. The succession beneath the Llanberis Slates and above the ash-flow tuffs is much thicker at Nantlle and includes the coarse grained Cilgwyn Conglomerate and the sand-rich Tryfan Grit Group and Glog Grit Group.

An area of well exposed Cambrian strata to the northwest of Llanberis was examined as an example of Arfon Group sediments on the Padarn Ridge (Fig 6.7).

FIG 6.5 Geological Succession: Llanberis (Padarn Ridge).

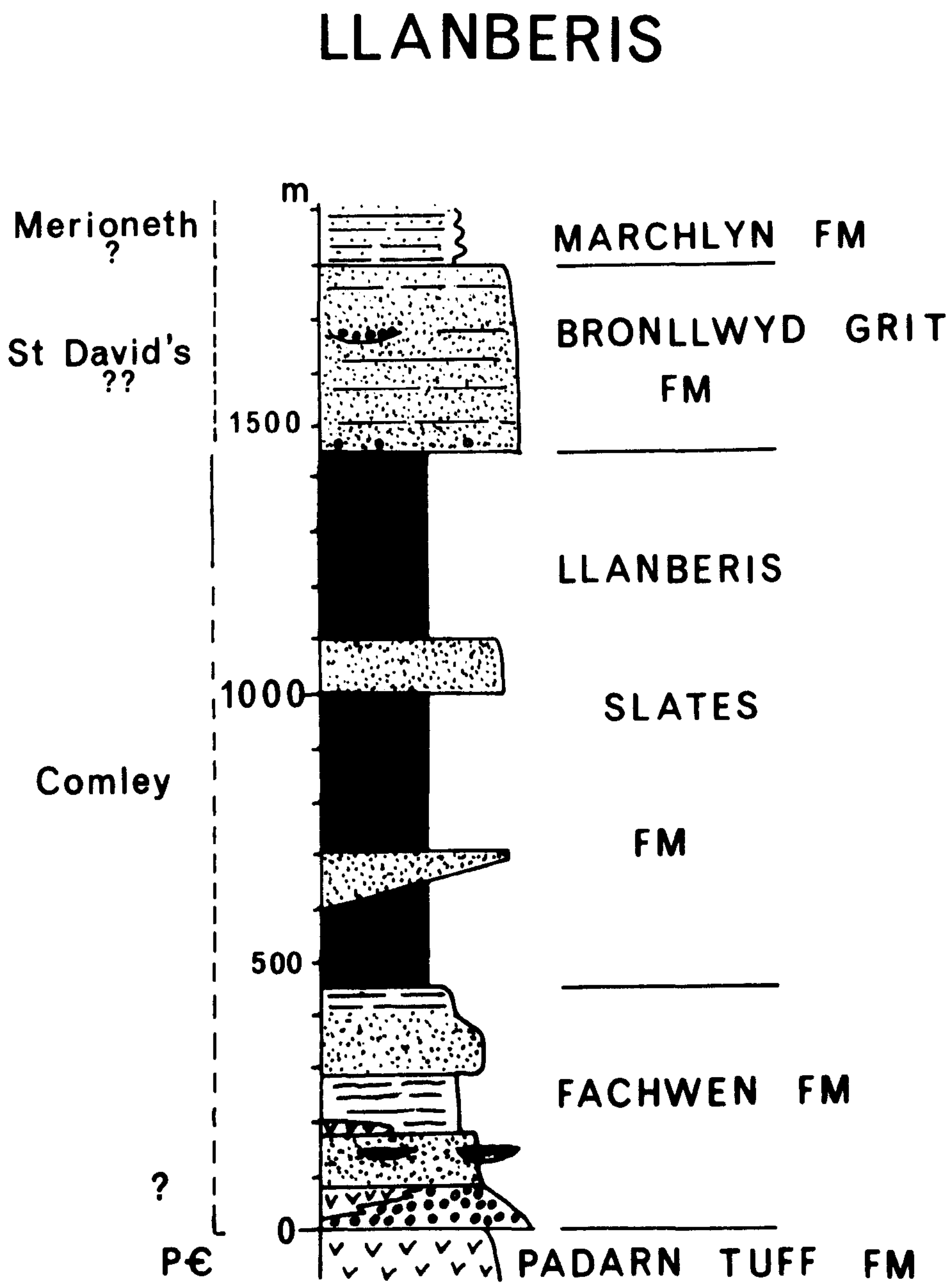
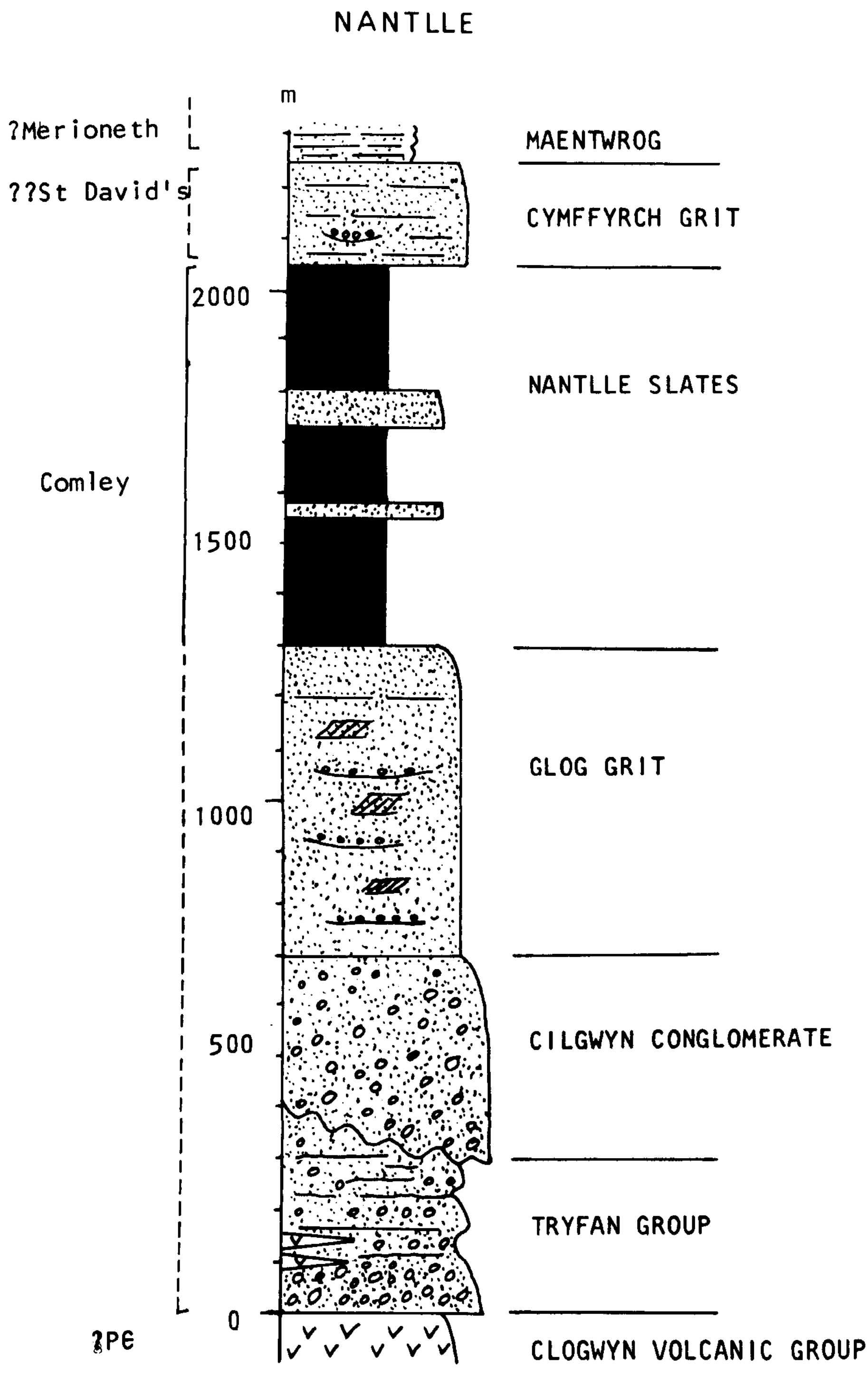
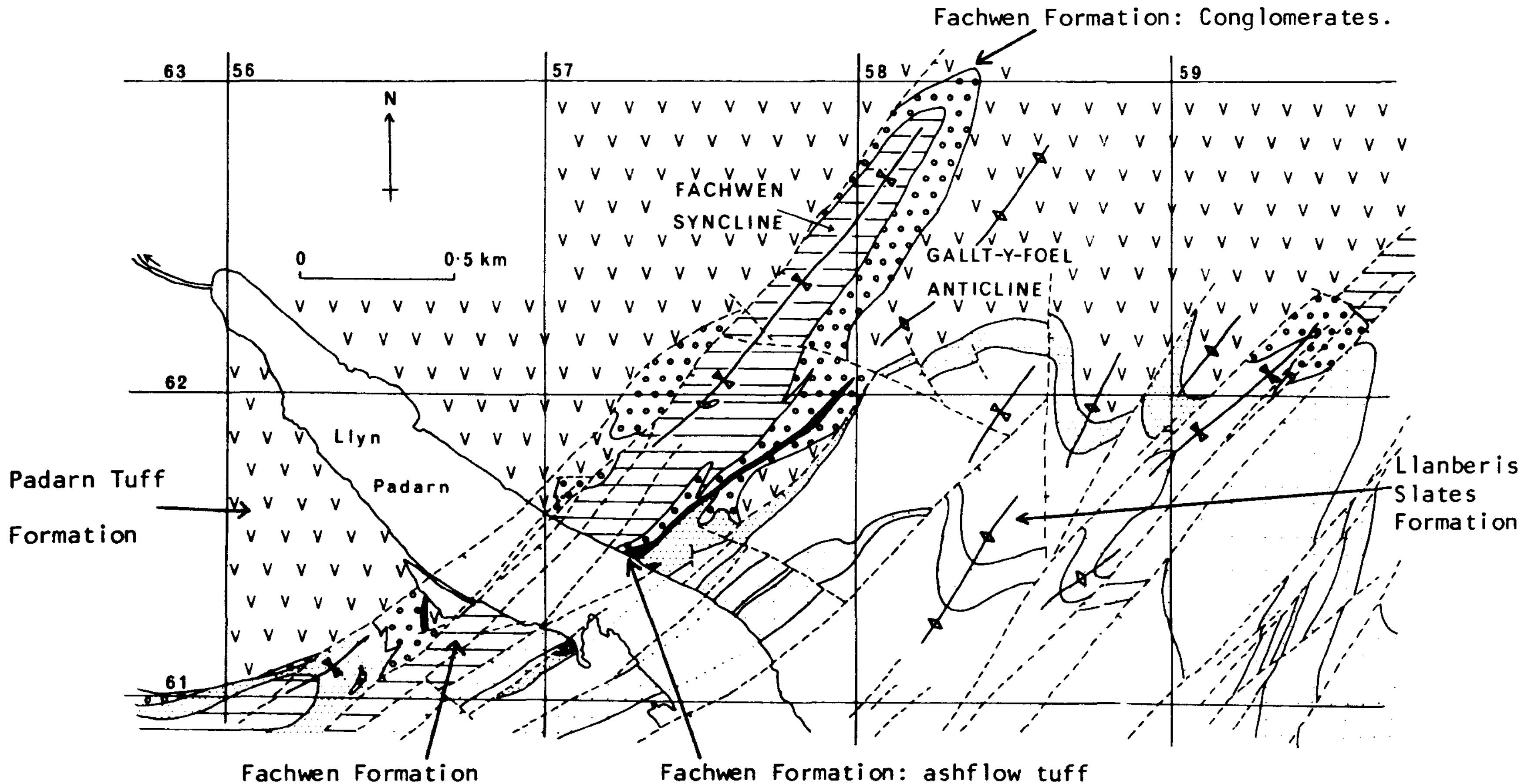


FIG 6.6 Geological Succession: Nantlle (Padarn Ridge).



(after Morris and Fearnsides 1926)

FIG 6.7 Geological Map of the north-east side of Llyn Padarn near Llanberis, Padarn Ridge.



(from B.G.S 1:25,000 Parts of Sheets SH65 and SH66)

1) Padarn Tuff Formation.

The Padarn Tuff Formation outcrops in the northwest part of Llyn Padarn (Fig 6.7) and is the oldest formation of the Arfon Group. Lithologically the tuff is relatively homogenous and massive and contains euhedral, skeletal and often cuspate crystals of quartz and feldspar. These megacrysts generally are 1-2mm in diameter and occur within a very fine grained, altered matrix. Pumice and lithic clasts are usually uncommon within the Padarn tuff, though at Cwm-y-Glo [SH 555 623] irregularly banded, darker lenses about 20-40mm long by 1-2mm thick occur and are probably squashed pumice fragments. At Cwm-y-Glo two varieties of tuff are exposed: a grey variety and pinkish-brown variety; the latter colour is probably due to post-depositional iron staining. The grey variety is generally most common and normally weathers to a white or light green colour. Shards are sometimes present and may show considerable flattening, though the amount of welding is very variable e.g. welding is mainly absent at Cwm-y-Glo while it is strong on the western side of Bigil [SH 577 622]. Silicification may occur parallel to the flattening fabric. The Survey (BGS, 1:50,000, Sheet 106) have used this welding foliation to determine the structure within the otherwise homogenous tuff and have concluded that the Padarn Tuff Formation is greater than 800m thick in this area. On the basis of gravity data (Reedman *et al.* 1984) calculated that the Padarn Tuff Formation is approximately 2km thick.

In some places the Padarn Tuff is brecciated e.g. north-west of Lake View Hotel [SH 562 617]. The clasts may be up to 100cm by 60cm, though usually they have a diameter of the order of 1-10cm. The clasts are usually of similar lithology to the matrix, though the former may be richer in megacrysts and the latter is commonly highly cleaved. Some of the brecciated bands are aligned parallel to the cleavage and clasts tend to be elongated parallel to this. The cleavage diverges around the clasts. It is not clear whether brecciation occurred during the deposition of the tuff or whether the breccia was tectonically produced. If the former is the case then the cleavage responded to the original inhomogeneity of the deposit. However nearby angular blocks of tuff occur in a tuffaceous matrix and are not aligned parallel with the cleavage and some of the blocks probably originally fitted together. This may be due to flow brecciation during the last stages of deposition of the tuff.

Northwest of the Lake View Hotel [SH 563 612] a 15-20cm thick bed is parallel laminated with laminae about 2mm thick. Some cross lamination also occurs with set thicknesses of the order of 15mm. Crystals of quartz and feldspar, volcanic fragments and red siltstone clasts tend to be concentrated within certain laminae. Since some of the laminae appear be graded and are rich in volcanic-derived clasts this bed was probably deposited as an air-fall tuff, composed predominantly of volcanic ash, but including segregated crystals and grains of epiclastic material.

The presence of shards, skeletal and cuspate crystals, welding and rare columnar jointing suggest that the Padarn Tuff Formation was mainly deposited from an ash-flow tuff with rare ash-fall tuffs. The general homogeneity of the deposit and absence of identifiable flow units suggest that the tuff was probably erupted over a short period of time and may have formed from ponded flows. Reedman *et al.* (1984) suggested that large scale pyroclastic flows, ponded within a volcanically produced graben, may have been produced by caldera collapse (c.f. Smith & Bailey 1968).

2) Fachwen Formation.

The Fachwen Formation is a unit of mixed lithologies which are to some extent transitional between the underlying volcanic dominated Padarn Tuff Formation and the overlying fine grained, clastic dominated Llanberis Slates Formation. The Fachwen Formation outcrops to the southeast of the NE-SW trending Aber-Dinlle Fault and is particularly well exposed in the Fachwen Syncline northeast of Llyn Padarn (Fig 6.7). The lower contact in this area is both conformable and unconformable on Padarn Tuff at different localities (see below) and the upper contact is poorly exposed but appears to be conformable and transitional.

The Fachwen Formation can be divided into the following facies:

A) Conglomeratic Facies.

This facies is characterised by very thick, mainly metre-bedded, disorganised, predominantly clast supported pebble conglomerates. Some matrix supported intervals also occur which usually contain pebbles in a matrix of medium sand. Diffuse alternations locally occur of pebble concentrated and pebble dispersed layers. A very fine grained, tuffaceous matrix, similar to the matrix between the megacrysts in the Padarn Tuff Formation is also very common in the basal conglomerates of the Fachwen Formation though higher up in the succession red to purple silt matrix is more common.

On the western limb of the Fachwen Syncline [SH 575 622] the Padarn Tuff Formation appears to pass up gradationally into Fachwen Formation conglomerates. The basal Fachwen conglomerates are similar to the Padarn Tuff apart from occasional pebbles, a feature observed by Ramsay (1866) and Wood (1969). The conglomerates get progressively more clast-rich upwards in a sequence approximately 2m thick. This sequence shows welded tuffs overlain by pebbly vitroclastic sandstone, the matrix of which looks like tuff. The contact must be abrupt (if pyroclastic deposits are overlain by waterlain deposits) but in the field the contact is indistinct. Reedman *et al.* (1984) have shown that in other parts of the Llanberis area there is sharp boundary and an angular discordance between welding fabrics (equivalent to bedding) in the Padarn Tuff Formation and bedding in the conglomerates. This indicates tilting and erosion of the Tuff in these areas prior to the onset of Fachwen Formation deposition and it is likely that some of the upthrown blocks supplied sediment to the areas where the base of the formation is conformable. There was probably therefore considerable lateral variability in topography within the Arfon Basin in early Fachwen Formation times.

Conglomerate beds are mainly massive. Some conglomerate beds are crudely graded and often have diffuse tops and bases (usually relating to different proportions of matrix). These features, together with the occurrence of some angular clasts indicate that the conglomerates were deposited by high density flows.

Where vertical grain size differences are abrupt this is often associated with scours, which occur at the base of some conglomerate beds. For instance Fig 6.8 shows an erosively based conglomerate eroding into parallel laminated and bedded sandstones which contain frequent pebble-rich laminae. Interdigitation of conglomerate and sandstone (Fig 6.9) at the base of some conglomerate beds indicates multiple fill of scours, though only rarely are there any major grain size differences within the conglomerate bed. Beds may show lateral variability on the larger scale; at outcrop (Fig 6.10) a thick, massive conglomerate lenses out laterally into several sandstone units, also suggesting multiple fill. Similarly massive conglomerates may pass laterally into laminated and bedded conglomerates. Therefore, although many conglomerates are thickly bedded and massive, they were not necessarily produced by single events, but by a succession of slightly higher and lower energy flows.

FIG 6.8 Scour filled with conglomerate, Fachwen Formation, Bigil.

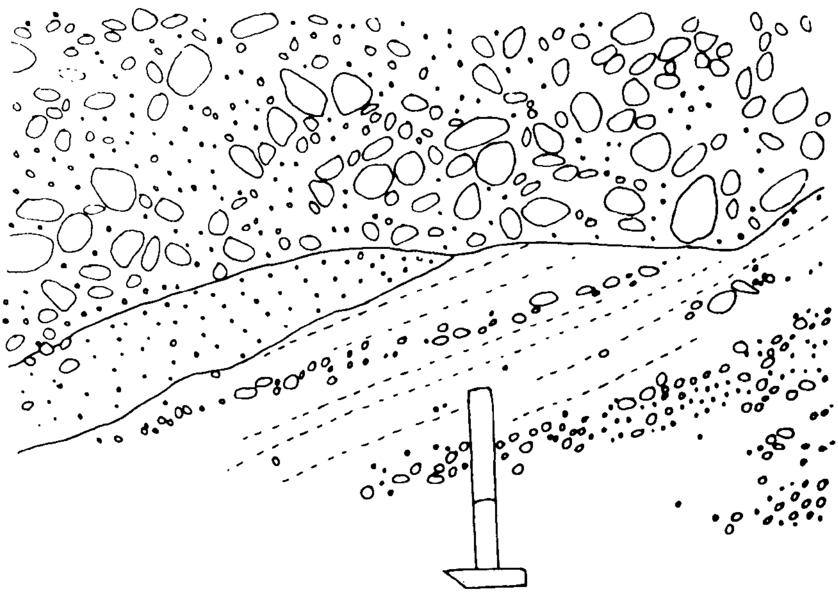


FIG 6.9 Lensoid conglomerates, Fachwen Formation, Bigil.

FIG 6.10 Lensoid top, conglomerates, Fachwen Formation, Bigil.

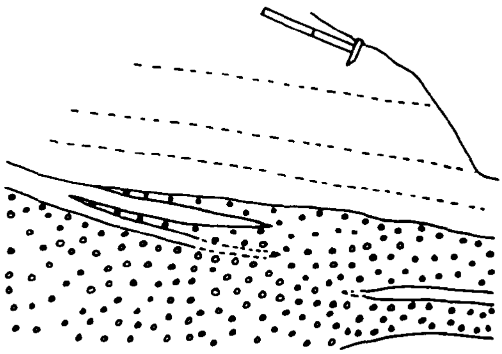
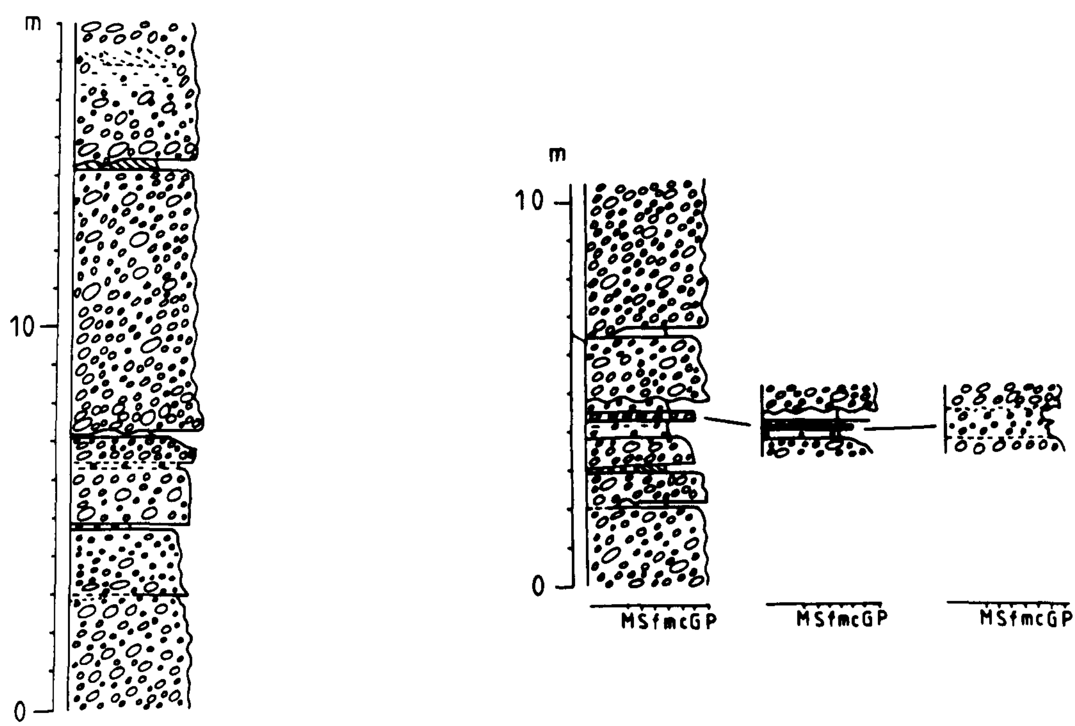


FIG 6.11 Logs in Conglomeratic Facies, Fachwen Formation, near Lake View Hotel.

Thick beds of conglomerate with thin sandstone interbeds.



South of the Lake View Hotel [SH 566 611] there are larger scale lateral variations; here thick conglomerates appear to pass laterally into thick bedded coarse sandstones interbedded with some conglomerates, so there appears to be interdigitation over several metres between the two facies. These lateral changes suggest possible channelisation of coarser grained deposits.

The conglomerates occasionally show evidence for traction including cross-bedding with set thicknesses of the order of several tens of centimetres (Plate 6/I). Faint laminae in the conglomerate dip shallowly towards the east and southeast and suggest the presence of large barforms of roughly 1m amplitude. The lack of clearly segregated laminae is indicative of rapid sedimentation during deposition.

The conglomerates are often interbedded with sandstones (largely medium grained) and only here is bedding well defined. The sandstones may form units a few centimetres to several tens of centimetres thick. However sandstone lenses up to 20cm thick are most common and are usually lensoid on the outcrop scale. The sandstones are usually well sorted and may be poorly laminated. However it is more common for them to be clearly laminated and occasionally cross-bedded as single sets of the order of 10cm thick and indicating flow towards the southeast (Plate 6/II). In contrast with the conglomerates the laminae are often very clearly defined and in some cases (e.g. Lake View Hotel [SH 566 612]) heavy mineral-rich (magnetite). The finer grain size and better laminated nature of many of the sandstones may indicate lower energy flows and slightly slower sedimentation rates than the conglomerates.

Few consistent vertical trends can be detected in logs taken from the conglomeratic facies (Fig 6.11), though many conglomerate beds pass upwards and laterally into sandstones.

The basal conglomeratic facies is particularly rich in volcanic clasts. Although these clasts vary in the proportion of quartz to feldspar megacrysts and the degree of welding, the majority are of Padarn Tuff type. White/grey

PLATE 6/I : Large scale cross-bedding, Conglomeratic Facies, basal Fachwen Formation, Bigil.

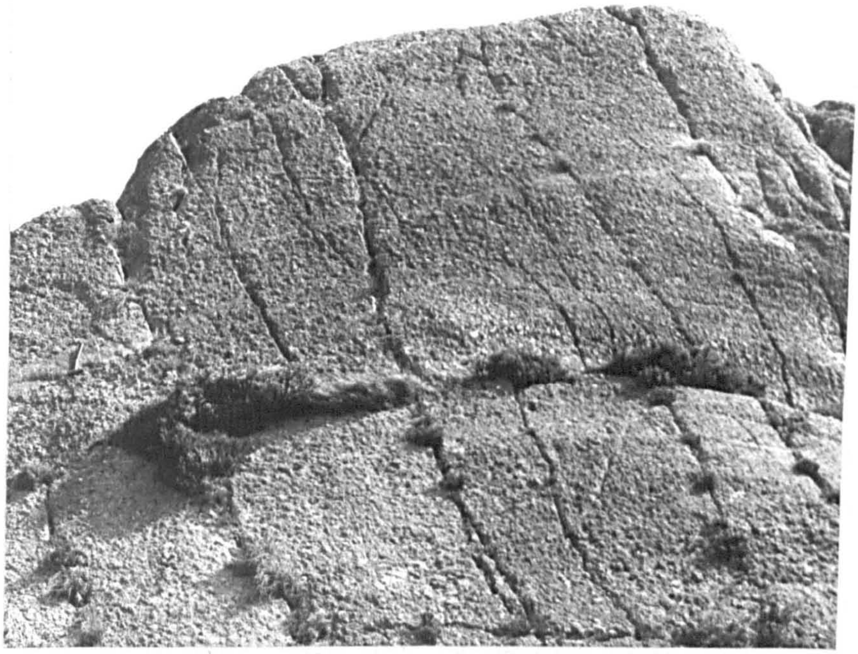


PLATE 6/II : Cross-bedded sandstones in Conglomeratic Facies, basal Fachwen Formation, Bigil.

varieties (Llanberis-type) are common, though some types are pinkish and others are dark brown (?Bangor-type Padarn Tuff) in colour. Other volcanic grains are also common including chert, intermediate composition ashflow tuff, siliceous dust tuffs (sometimes parallel laminated), flow banded rhyolite and occasional dark basic volcanic clasts. Reedman *et al.* (1984) observed a decrease in the proportion of acidic to more basic volcanic clasts upwards in the Fachwen Formation and this is accompanied by a reduction in the proportion of clasts of Padarn Tuff type. In addition the conglomerates also contain acidic plutonic pebbles mainly of microgranite (both equigranular and porphyritic) and include rare grains showing micrographic texture. Some clasts are of similar lithology to the Twt Hill granite which outcrops at Caernarfon and intrudes, though may be broadly coeval with the Padarn Tuff (Greenly 1944). However the granite has been dated at 498 ±7 Ma (Thorpe *et al.* 1984), considerably younger than the Lower Cambrian conglomerates. If this is a magmatic age then the Twt Hill granite could not have supplied the Fachwen Formation with detritus. However if this is a reset age then it is possible that the granite was intruded prior to deposition of the Fachwen Formation. Sedimentary quartzite and some siltstone grains are also present. Isolated grains of quartz and less commonly feldspar also occur, usually of a similar size to the megacrysts in the Padarn Tuff. Therefore the conglomerates are mainly of volcanic origin reflected in both pebble type and matrix composition. A large proportion of the clasts are locally derived, though some were supplied from further afield (more basic volcanic clasts, microgranite, sedimentary clasts; some of the quartzite clasts may have come from a metamorphic source).

The massive, disorganised, often poorly sorted nature of the conglomerates suggests deposition from relatively large, dense flows possibly on alluvial fans. However the rare occurrence of large scale cross-bedding, produced by bar bedforms in a facies directly overlying and grading up from subaerial ashflow tuffs indicates that deposition occurred within a gravel dominated, braided fluvial system. The conglomerates probably indicate rapid deposition from high stage flows while the interbedded sandstones represent deposition at times of low stage. The sandstones have a lensoid geometry either due to scour of originally tabular sand beds followed by deposition of conglomerate at high stage or as a result of restriction of flow to certain channels during periods of low stage.

Regionally the tilting, uplift and erosion of some blocks at the same time as the subsidence of adjacent blocks (where sedimentation was continuous) probably resulted from normal faulting within a tensional regime. Fault movements probably controlled the location and geometries of the alluvial fans and influenced the lateral variability of this facies in the Arfon area. The general decrease in the abundance of clasts of Padarn tuff type and the progressive increase in more basic clasts upwards in the Fachwen Formation may indicate either:

a) At the base of the Fachwen Formation local areas of subsidence controlled sedimentation, whereas higher up in the sequence regional subsidence in Arfon predominated, being fed from more distant rather than local sources. Although many of the clasts are well rounded some (mainly volcanic) are highly angular which would indicate local derivation.

b) There was gradual unroofing of the Anglesey area; at first volcanic overspill from the Arfon Basin was eroded, then basic volcanics from the Mona Complex. This is a view favoured by Reedman *et al.* (1984).

It is possible that subsidence in the Arfon area in Fachwen Formation times resulted from caldera collapse after the large scale eruption of the Padarn Tuff (Smith & Bailey 1968; Reedman *et al.* 1984). The deposition of tuff would have been closely followed by deposition of conglomerate, deposited in alluvial fans and braided fluvial environments and derived from the margins of the caldera.

2) Cross-bedded Sandstone Facies.

This facies most commonly occurs above the Conglomeratic Facies, near the base of the Fachwen Formation. The beds are commonly composed of medium to very coarse sandstone. Larger grains (up to pebble size) occur more rarely, sometimes as diffuse pebble trains. Individual laminae are well sorted and there is much less fine grained matrix than in much of the Conglomeratic Facies. These sandstones are typically trough cross-bedded (e.g. at Lake View Hotel) with set thicknesses of 7-20cm and a mean of about 15cm. Cosets are present and flow was probably mainly towards the southwest.

On the west side of Bigil [SH 574 621] this facies is also present. Here beds are on average about 10cm thick (range 4-30cm) though bed boundaries are generally diffuse, with an alternation of very coarse sandstone to granule conglomerate and coarse to medium sandstone. Beds are discontinuous laterally and seem to occupy broad, shallow scours, often associated with faint trough cross-bedding/lamination, with mean set amplitudes commonly of about 8cm. Low angle cross lamination is much more common at Bigil than at the Lake View Hotel exposure.

The well sorted texture and abundance of well rounded clasts indicates that there was winnowing out of fine grained material prior to deposition of this facies. Since this facies is generally finer grained than the Conglomeratic Facies, which commonly underlies the Cross-bedded Sandstone Facies, the latter was probably deposited by lower energy flows than the former. Trough cross-bedding was produced by sinuous crested bedforms. The coarser grains may represent lag deposits which were only transported during occasional periods of higher flow velocity.

The Cross-bedded Sandstone Facies probably represents braided fluvial deposits, which were part of a sand-rich system more distal from the apex of the alluvial fan(s). It is possible that the Arfon basin was fed from marginal sources by conglomeritic alluvial fans (Facies 1) while sand-rich braided fluvial systems flowed axially (Facies 2).

3) Red/Green Siltstone Facies.

This facies is predominantly made up of thin bedded (less than 5cm thick), finely laminated green and red siltstones, often with graded laminae and include thin laminae of sand. Parallel lamination is common, though locally cross lamination and convolute lamination are present. This facies is most common in the upper part of the Fachwen Formation. Although on the large scale green and red siltstones tend to occupy different parts of the succession, in detail they may be interbedded with each other and may pass laterally from one coloured siltstone to another. This lateral change in colour suggests that although the larger scale differences may have been depositionally controlled, in detail the variability in colour was diagenetically controlled.

As well as the relatively monotonous alternations of siltstones, sand laminae and mudstones, other bed types also occur in this facies:

a) Conglomerates. Conglomerates may be clast supported but are more commonly matrix supported; the matrix is usually composed of red silt. The clasts are usually of granule to pebble size. These beds often have lensoid geometries (see Wood 1969) and contain clasts of red tuff (some with phenocrysts), altered fiamme and occasional grains of jasper, microgranite, chlorite and different types of tuff. South of Bigil mast [SH 575 620] a thick conglomeratic bed outcrops. It is several metres thick, has an undulose, loaded base and has intruded coarse to medium sand as sedimentary dykes into the underlying green (and locally reddened) siltstones (Plate 6/III). This conglomerate contains both acidic volcanic clasts of Padarn Tuff type (some of which are haematite stained) and darker, more basic volcanic clasts. In general the basic fragments

PLATE 6/III : Sedimentary dykes from conglomerate bed, Fachwen Formation, Bigil.

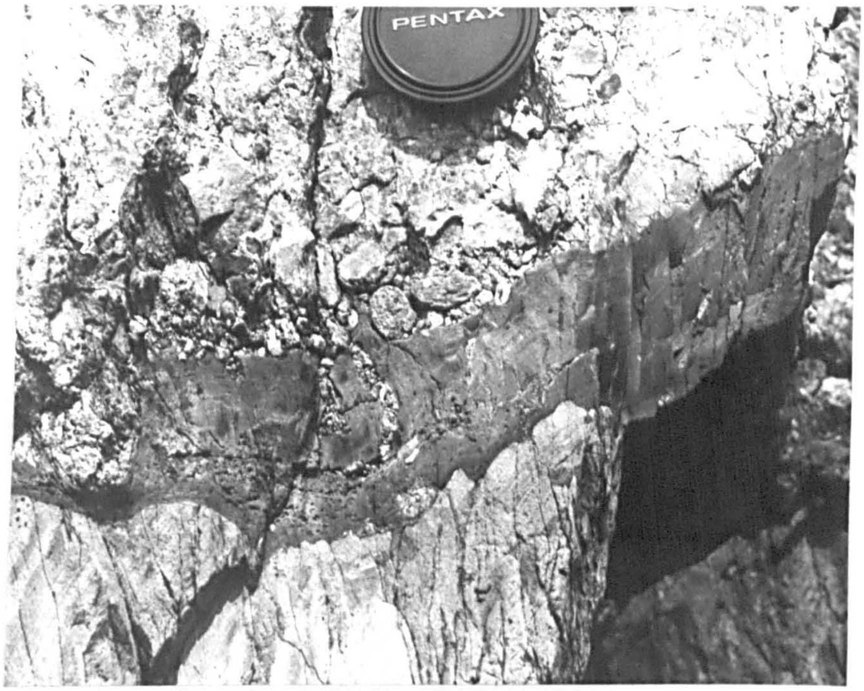


PLATE 6/IV : Lensoid beds and laminae in Sandstone-siltstone Facies, Fachwen Formation, Bigil.

are relatively well rounded and together with occasional jasper clasts may indicate a Monian type source. The acidic clasts probably indicate a more local supply of sediment.

b) Sandstones. The sandstones are often parallel laminated, though some are cross-bedded (up to 40cm set thickness) or convolute laminated. Some beds show graded bases and are diffusely laminated and may be rich in intraclasts.

c) Green Tuffaceous Beds. These beds are generally fine grained and often hard and siliceous.

Abundant graded laminae indicate that this facies was largely deposited from dilute flows, possibly in a relatively low energy shallow marine or lacustrine environment. The high proportion of volcanically derived clasts indicates a volcaniclastic origin and some beds/laminae may represent thin ash-fall tuffs or redeposited tuffaceous material. Massive bedding and a high proportion of matrix in conglomerate beds indicate occasional deposition from higher density flows.

4) Sandstone-Siltstone Facies.

Facies 4 is transitional with facies 3 and differs only in the increased proportion of sand within the facies. The facies is usually thinly bedded, beds rarely exceeding 10cm thick and are often finely laminated, showing a strong bimodality in grain size between fine to medium sand and silt. Many beds and laminae are lensoid (Plate 6/IV) and sand intrusions are also very common. Lateral variations from massive sandstones to laminated sandstones are also common and many of the sandstones are highly feldspathic. Some upper boundaries of laminae are abrupt while others are graded; some beds show gradual thinning upwards of laminae above thicker sandstone beds as part of sequences usually less than 5cm thick.

Facies 4 was probably deposited in a similar environment to facies 3, except that it was richer in sand.

5) Sandstone Facies.

This facies is predominantly made up of beds of sandstone which are commonly 10-30cm thick and up to 80cm thick. These sandstones may be amalgamated in packets or interbedded with thin bedded siltstones. The sandstones are often relatively coarse grained and occasional granule conglomerates may be present.

The following bed types are present in this facies:

a) Relatively massive, faintly laminated fine to coarse sandstones.

b) Parallel laminated sandstones.

c) Parallel laminated siltstones.

d) Low angle cross stratified fine to medium sandstones.

Most bed boundaries are discontinuous and only a few can be followed for the width of the outcrop (e.g. east of Bigil [SH 577 621], for 30-40m). Relatively massive beds pass laterally into separate beds of sandstone and siltstone. Other sandstones have a lensoid geometry (wide and shallow) but there is no preferential direction of thickening or thinning. Rare lenses (about 10cm thick) of intraclast breccia also occur.

Cross-bedding is locally common. Set heights of about 10-20cm are the most common and both tabular cross-bedding and low angle cross-bedding occur. Laminae often thicken and thin over topography on bedset boundaries and occasionally drape erosional surfaces. Internal discontinuities also occur between laminae. Some sets have well defined foresets with alternating laminae of coarse to medium sandstone and red siltstones. Tabular and trough cross-bedding occur in this facies indicating straight and sinuous crested ripples. However some beds have drapes and internal discontinuities between laminae, similar to hummocky cross stratification described from storm deposited sands which have been affected by storm wave activity (Dott & Bourgeois 1980; Brenchley 1985).

Lenticular sand beds/laminae, cross-bedding and possible hummocky stratification may indicate deposition in shallow water, occasionally affected by storm driven currents.

6) Tuffaceous Facies.

Tuffs occur at several horizons within the Fachwen Formation. In a cutting on the Padarn Lake Railway (north-east shore of Llyn Padarn) [SH 5728 6147] tuffs are particularly well exposed in a 15m thick sequence. It is difficult to determine their exact stratigraphical position, though they appear to overlie basal conglomerates and cross-bedded sandstones of facies 1 and 2. The sequence can be divided into the following units:

A) Over 1.5m of grey-green massive crystal tuff. It is welded, showing flattening of some grains around 1-2mm long quartz phenocrysts. Chlorite-rich clasts up to 3mm long occur and may represent altered fiamme or lithic fragments. Occasional clasts of welded tuff are also present. The matrix is composed of fine grained sericite and chlorite and is proportionately more important than the phenocrysts and clasts.

B) 2.4m of green massive crystal tuff. This also contains 1mm diameter phenocrysts of quartz, some of which show skeletal structures. This rock type is generally more massive and glassy than unit 1 and is poorer in chloritic clasts. This unit also contains smaller grains (less than 1mm diameter) of brown/white weathered, parallel banded rhyolitic vitric tuff.

C) 8m of green and red lapilli tuffs. This unit has a sharp base and an abrupt, eroded top. The tuffs contain abundant volcanically derived clasts which have a maximum length of 20cm (lapilli) and a mean length of the order of 0.5-1.0cm, though grains up to 2cm in length are common. The tuffs are poorly sorted and matrix rich, with a matrix of fine grained green and red-brown volcanic ash. This unit lacks any clear coarsening or fining upwards trends, instead larger clasts are distributed irregularly as outsize blocks within the poorly sorted deposit. The unit is massive and lacks internal organisation.

The lapilli tuffs are strongly flattened, the matrix being composed of flattened ash fragments, probably of intermediate composition. The following clasts types are present: fine grained light green and red tuffs, dark green chloritic fragments, black tuff with altered ?feldspar phenocrysts, welded crystal tuffs, small dark shardic fragments which are locally flattened and preserve their Y-shape, cuspate or lunate outlines, and rare micrographic quartz-feldspar grains. Many of the grains are irregular and angular (Plate 6/V), sometimes with re-entrant, concave-outward outlines. Therefore the grains have suffered little abrasion during transport. Some grains have a 1-2cm thick white ashy band surrounding them, often resulting in a gradational boundary with the matrix which suggests that the outer part of the clast may have reacted with fluids and gases during transport or deposition (Plate 6/VI).

Most clasts are elongate in section, due to their disc-like or tubular shape, though rarer large clasts appear to be more blocky in shape. Many clasts have relatively smooth outlines parallel to the long axis of the grain while their outlines perpendicular to this are often jagged; this probably results from the flattening of pumice and chloritic volcanic clasts. Most clasts have a length-width ratio (when viewed on vertical surfaces perpendicular to the cleavage) of between 1:4 and 1:10. The larger grains in general have a lower ratio than the smaller grains. The clasts are aligned parallel with the cleavage, which dips to the northwest much more steeply than bedding. At the northwest part of the exposure matrix-poor conglomerate (unit D) with mainly rounded clasts abruptly overlies unit C. The cleavage and grain alignment in unit C is discordant with the bedding, which dips gently towards the northwest. Unit D is poorly cleaved and thus the alignment of the clasts in unit C can be seen to have been tectonically produced. Cleavage would have been produced as a result of compression, though it is

PLATE 6/V : Cuspate grains in ash-flow tuff, Fachwen Formation, Llanberis Lake Railway cutting.

PLATE 6/VI : Lapilli in ash-flow tuff, Fachwen Formation, Llanberis Lake Railway cutting.

unclear whether all the flattening of the clasts was tectonic, or whether some flattening could have occurred during welding before the clasts were rotated in to parallelism with the cleavage.

Some grains (notably the quartz phenocrysts) have resisted flattening and other grains have flattened around them. The rotation of the grains within the matrix must have been entirely passive since there is little evidence of breakage of even the most delicate chloritic clasts. The penetration of cleavage into the overlying conglomerate (unit D) suggests that the cleavage is not a welding foliation that has since been tilted. Therefore the contact is not an unconformable one as suggested by Blake (1893).

D) Over 3m of dark red-brown conglomerate. The basal erosive contact is loaded and is locally silicified. The conglomerate is composed mainly of rounded small and large pebbles. The clasts are predominantly composed of dark green and grey volcanic fine grained sandstone with occasional chert clasts. This unit is much better sorted than unit C.

The massive nature of the tuffs and the presence of welding suggest that the tuffs were probably deposited from a pyroclastic flow. The deposits are silicic and are similar in some ways to the Padarn Tuff Formation which probably occurs a few tens of metres below this horizon.

Unit C is a lapilli tuff and is probably more basic in composition. The presence of volcanic lithics and shards suggest that it is of volcanic origin. Its massive nature and the lack of reworking indicate that that it was rapidly emplaced. The angularity and delicate nature of many of the clasts suggests that they were produced by explosive volcanism but have suffered little abrasion during transport. The lack of sorting and normal grading indicate that it probably was not an airfall deposit, whereas the massive nature of the deposit and the possible occurrence of welding suggest deposition from a pyroclastic flow probably in a subaerial environment. Unit C is more lithic and pumice rich, is more basic in composition than units A and B and is erosively overlain by probable fluviatile conglomerates eroded from areas of renewed tectonic uplift.

The ignimbrite (unit C) can be traced for several hundred metres laterally within the Fachwen area. However larger scale lateral correlation is more difficult since exposure is often poor and facies are highly variable in thickness. However it seems likely that it can be correlated with the following localities:

i) South of Bryn Bigil [SH 5783 6190]. Here the tuff is about 5m thick. The tuff is fine grained and locally silicified with feldspar phenocrysts and nodular green and red patches, 2-3mm in diameter. In some places the nodules coalesce and some have been rotated and/or compressed into the cleavage. The tuff also contains irregular lenses up to 7cm long and 1.2cm thick infilled with red-purple chert, some of which is parallel laminated. Many have irregular stylolitic-like margins and laminae appear to thicken into the central part of the lense. These lenses may be silicified volcanic intraclasts.

ii) North of Wylfa [SH 5810 6243]. This locality is similar to locality 1. The tuff contains nodules which are commonly 1-15mm long (maximum 30mm) some of which are silicified. The nodules in this part of the ash-flow may have formed by degassing of the pyroclastic flow as a result of contact with water (as suggested by Wood 1969).

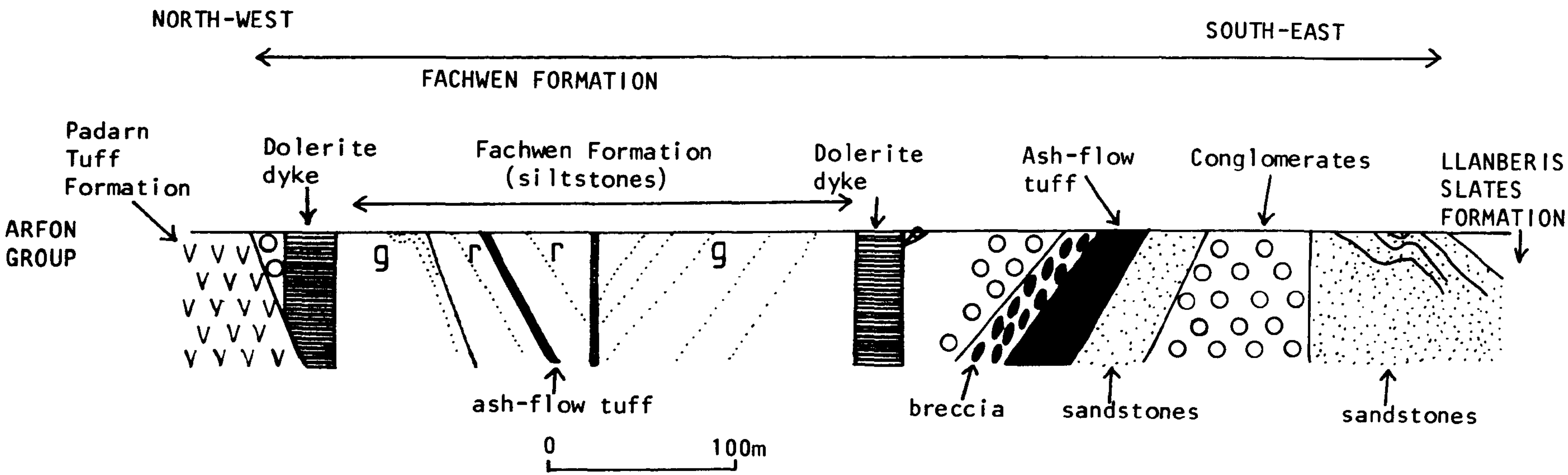
The ash-flow described above can be correlated along most of the eastern limb of the Fachwen syncline but seems to be largely absent on the western limb, except near the Lake View Hotel [SH 566 612].

Another ash-flow tuff can be recognised in a cutting on the Padarn Railway on the north-east side of LLyn Padarn (Fig 6.12). Here a tuff over 2m thick is exposed between thin bedded parallel laminated siltstones (facies 3). Lithologically it is similar to the Padarn Tuff and units 1 and 2 of the ashflow described above.

Interpretation.

There was continued volcanic activity after the eruption of the Padarn Tuff Formation. Volcanic products

FIG 6.12 Section along the Padarn Lake Railway.



(g: green, r: red)

included both subaerially erupted pyroclastic flows (facies 6), thin airfall and redeposited volcaniclastic deposits (facies 3 and 4). Overall the Fachwen Formation fines upwards in the following sequence:

a) alluvial fan, braided fluviatile conglomerates containing large bedforms, produced at high stage and smaller, sandy bedforms at lower stage.

b) braided fluvial sandstones often containing cross-bedding.

c) shallow marine siltstones and sandstones which were probably influenced by storm activity.

Thus the Fachwen Formation can be interpreted as a transgressive sequence from fluviatile deposits at the base to shallow marine deposits, passing up gradationally into the deeper water, basinal Llanberis Slates Formation.

3) <u>Llanberis Slates Formation.</u>

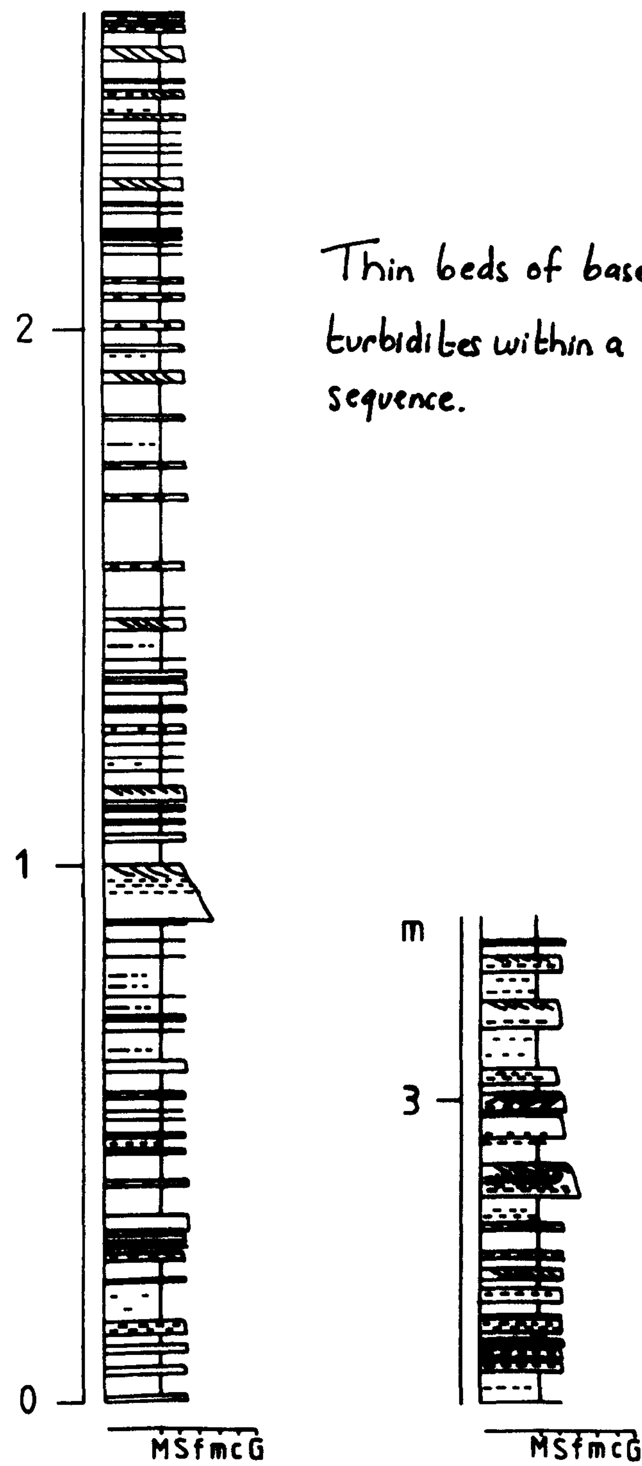
The Llanberis Slates Formation is largely comprised of cleaved purple, blue-grey and green silty mudstones. It forms a sequence 800-1000m thick (Cowie *et al.* 1972) and outcrops in a NE-SW aligned belt running from Bethesda through Llanberis to Nantlle.

Much of the Llanberis Slates Formation is composed of unlaminated or faintly laminated mudstones and siltstones with few sedimentary structures. However about 100m southeast of Cei Llydan Station on the Llanberis Lake Railway [SH 576 614] a more heterogeneous sequence is exposed (Fig 6.13). Here there is an alternation of dark green very fine sandstones and light green siltstones and mudstones. The dark green sandstones are typically thinly bedded ranging from 1-10cm thick. Some beds are graded and parallel laminated (c.f. Piper 1978) and thin Bouma T_{cde} and T_{de} sequences also occur. Rare thicker beds contain T_{bcde} sequences with occasional climbing ripples and convolute lamination. The light green siltstones and mudstones are usually well cleaved and contain faint, diffuse laminae. Therefore this sequence is similar to the Thin Bedded Facies of the Rhinog Formation and the Gamlan Formation and was probably also deposited from dilute turbidity currents.

The Llanberis Slates Formation also contain occasional thick sandstone beds which may be clustered within the sequence. For instance at Chwarel Fawr quarry, Dinorwic [SH 588 617] (Fig 6.14) 25cm to 2.5m thick beds typically contain T_{ae} and T_{abe} sequences interbedded with thinner beds. Above Dinorwic [SH 592 614] scours occur infilled by coarse grained, thick bedded sandstone. Multiple grading is also present (Plate 6/VII) and possibly indicates deposition from higher density flows and/or surging flow conditions.

Some of the sandy turbidite packets are laterally continuous within Arfon e.g. the Dorothea Grit of Nantlle

FIG 6.13 Log in the Llanberis Slates Formation, Llanberis Railway.



Thin beds of base absent turbidites within a mud-rich sequence.

FIG 6.14 Log in the Llanberis Slates Formation, Chwarel Fawr quarry, Dinorwic.

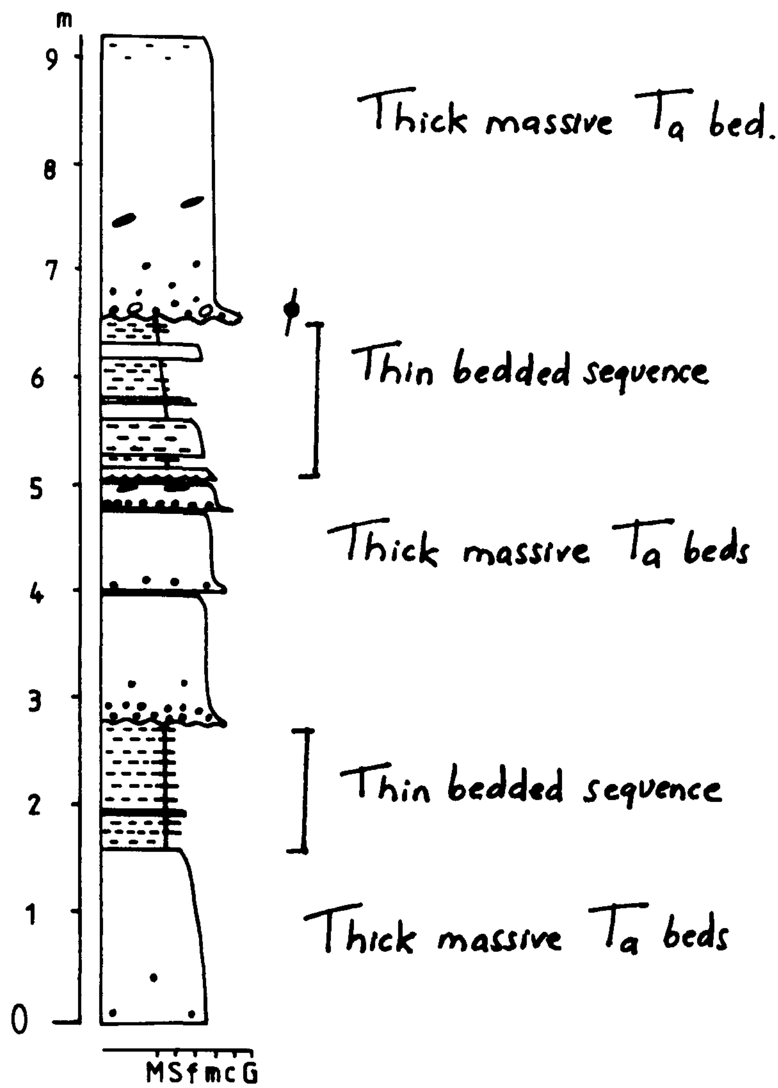
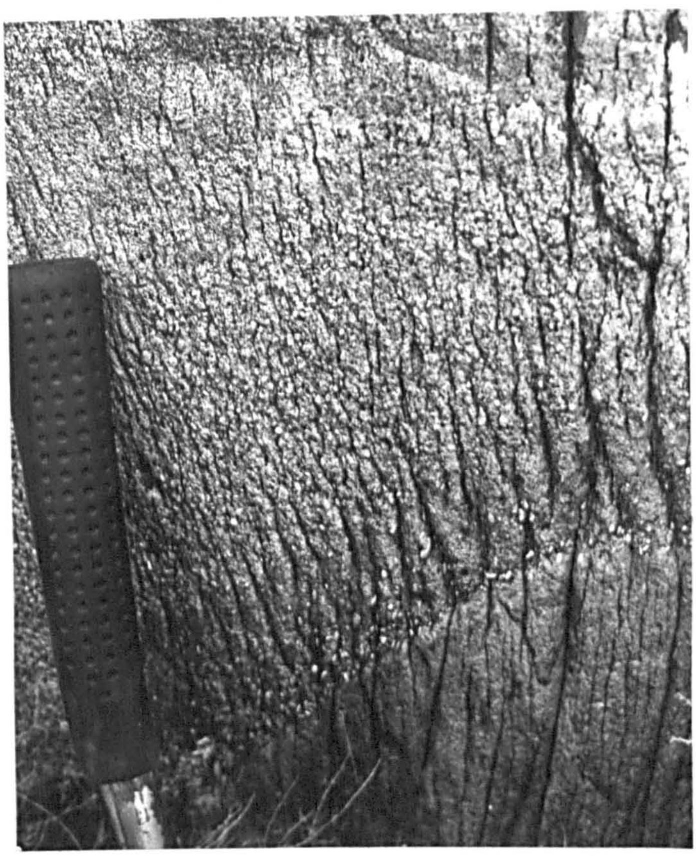


PLATE 6/VII : Multiple grading, sandstone packet in Llanberis Slates Formation, near Dinorwic.



can be correlated with the Lower Grit of Llanberis and the Pen-y-bryn Grit of Nantlle with the Dwndwr Grit of Llanberis. In general the grits get thicker towards the northeast (Cowie *et al.* 1972). However Webb (1983) identified thickness changes in the Dorothea Grit and interpreted these as a response to syn-sedimentary faulting, infilling small grabens and thinning over the horsts. Therefore overall the sandstone packets have a sheet-like geometry, though in detail faulting affected the thickness of sediment deposited.

The Llanberis Slates Formation was mainly deposited from dilute turbidity currents with occasional deposition from sandy turbidity currents, the sedimentation of which was fault controlled. The derivation of the sand turbidites is unclear since palaeocurrents are lacking, though it may have been from the northeast on the basis of sandstone thickness variations.

4) Bronllwyd Grit Formation.

The Bronllwyd Grit Formation occurs in the southeast part of the Padarn inlier and is predominantly composed of sand-rich turbidites. The formation is about 500m thick in the Llanberis area; it has a sharp erosive base but a relatively gradational top, being overlain by the Marchlyn Formation. The age of the Bronllwyd Grit Formation and its correlation with other areas is problematical (see chapter 3), though it may be late Lower Cambrian and thus correlate broadly with the Rhinog Formation of the Harlech Dome.

In general beds are massive and thickly bedded. In the Llyn Peris section [SH 597 588] (Fig 6.15) the mean sandstone bed thickness is 76cm with a range of 3-165cm. Sandstones generally dominate sequences (in the Llyn Peris section the sandstone-siltstone ratio is 13) and most beds are composed mainly of coarse or medium grained sandstone with some fine sandstone. Granule conglomerates occasionally occur and dispersed pebbles and larger intraclasts may be common locally. Bases of beds are often erosive and coarse sediment is often restricted to these shallow scours. These bed bases may also be loaded and include mud injection (flame) structures (Plate 6/VIII). Graded bedding is also very common. Coarse tail grading is most abundant and is usually restricted to the lowest part of a particular bed. Beds often do not grade upwards gradually throughout their thickness, instead they may grade up relatively abruptly in certain parts of the bed. In detail the grading may be interrupted by 5mm thick, discontinuous, finer laminae which usually have a spacing of about 2-3cm. They may be interpreted as shear laminae resulting from high density affects during sedimentation (Lowe 1982). Cross-bedding occasionally occurs and is generally more common near the base of the formation e.g. 8cm set height (south of Penrhyn Quarry, [SH 620 640]). Convolute lamination is also present rarely. Sequences are dominated by T_a Bouma divisions, both as the basal division and relative to the other

FIG 6.15 Log in the Bronllwyd Grit Formation, Llyn Peris.

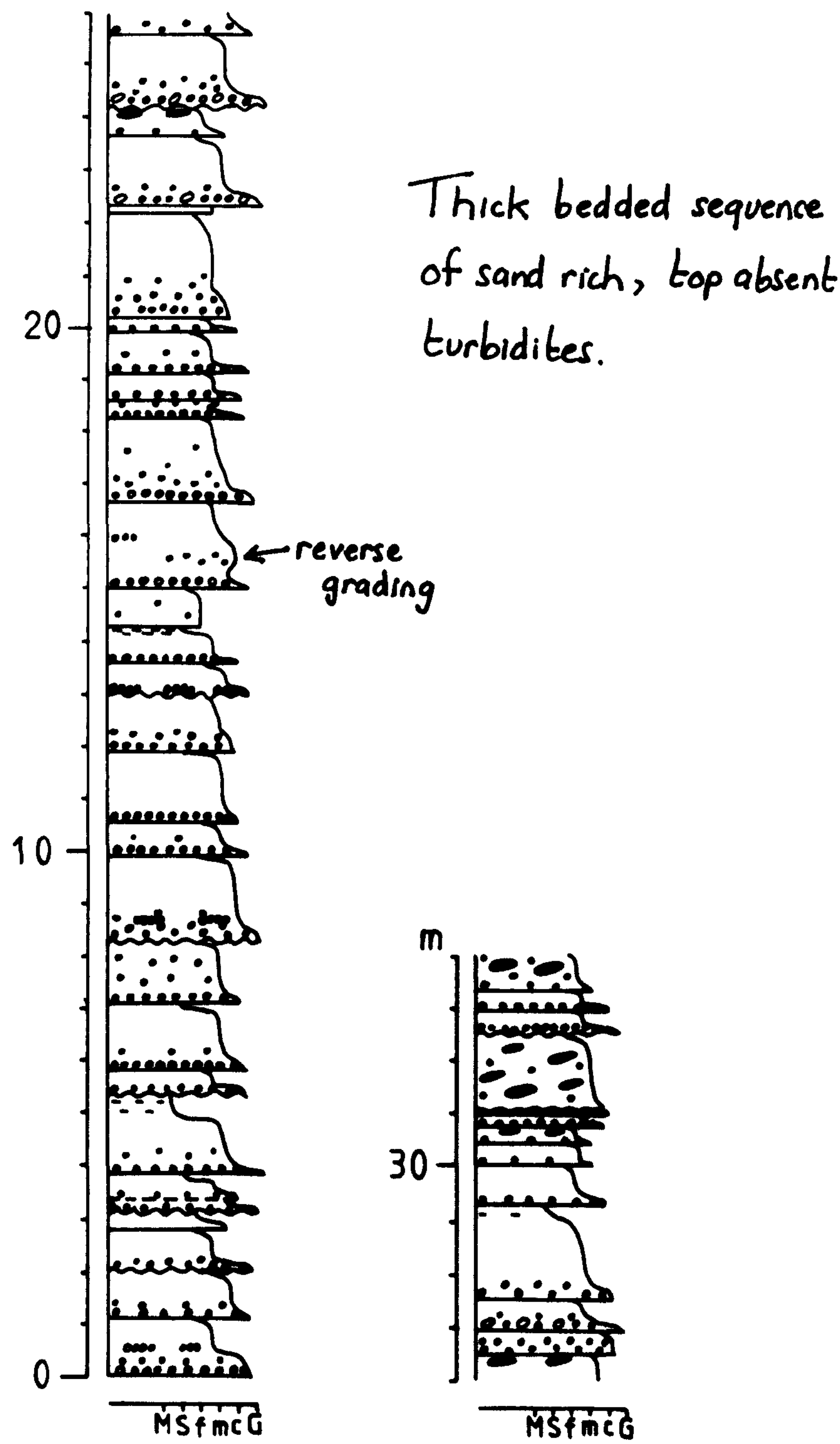
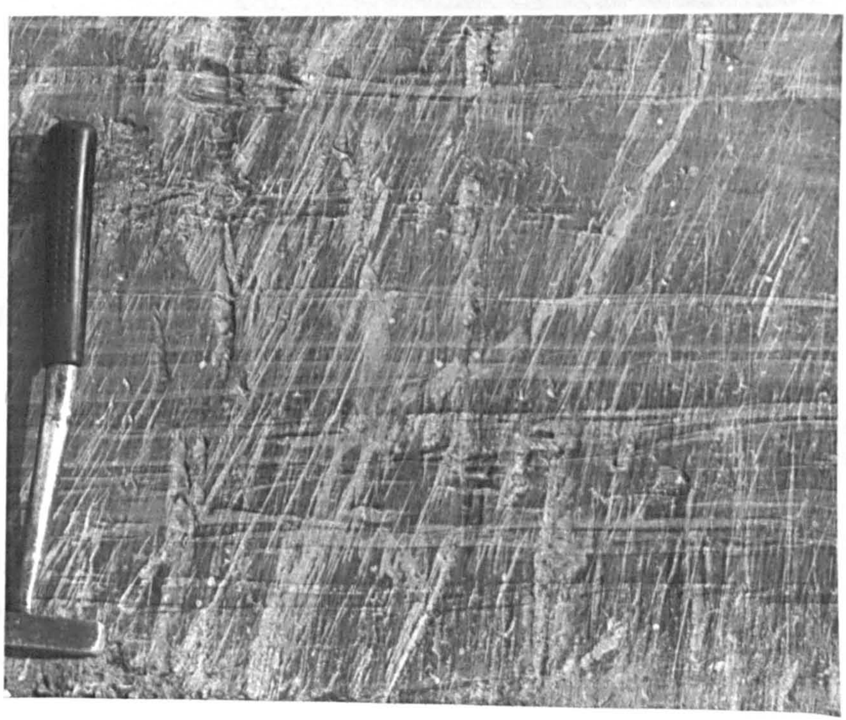


PLATE 6/VIII : Sole structure, Bronllwyd Grit Formation, Llyn Peris shore.

PLATE 6/IX : Thinner bedded units, Bronllwyd Grit Formation, Llyn Peris shore.



divisions. Occasional T_b divisions occur, usually as part of T_{ab} sequences, but amalgamated T_a sequences and T_{ad} are most common. Therefore the Bronllwyd Grit Formation is dominated by top and middle absent sequences.

Thinner bedded (Plate 6/IX), finer grained units are also present locally, e.g. exposed on the ice-scoured surfaces on the northeast shore of Llyn Peris [SH 598 592]. The thinner bedded units are usually interbedded with thicker sandstone beds and often contain abundant intraclasts, some of which have contorted laminae. Most of the thicker beds in this section are laterally continuous on the scale of the outcrop (30m). However the thinner beds (less than 10cm thick) may be grouped into a separate facies and form units up to a metre (usually several decimetres) thick and often show considerable lateral variation. Beds and/or laminae may range laterally from a few millimetres to several centimetres thick and may lense out completely laterally within a few metres. Lenses up to 20cm thick may occur and parallel laminated beds may pass laterally into more massive units. Folding is common and microfaulting may be present, much of which seems to be soft sedimentary in origin. These features may be associated with chaotic loading and sand injection structures. These beds often contain lenses of sand which may have been produced by loading into the underlying siltstone as ball and pillow structures. Thus much of the lateral variability may have resulted from heterogeneous soft sediment deformation.

It is possible that soft sediment deformation may also be associated with the formation of intraclasts in the interbedded thicker sandstones. In some places there appears to have been some post-depositional collapse of some of the thinner, finer grained beds down into the underlying thick bedded sand. Loss of strength within the sand may have resulted from liquefaction and resulted in failure of the finer grained "roof" to the sand beds followed by concomitant sand injection into the thin bedded facies. Taken to an extreme the failed "roof" might have broken up and possibly account for some angular intraclasts in some of the sandstones. Some beds contain dispersed coarse grains within a much finer cleaved matrix and may include abundant large intraclasts. It is possible that these represent the deposits of particularly high density flows. Some intraclasts contain folded laminae which may result from shear within these flows.

In summary, the Bronllwyd Grit Formation is dominated by thick bedded, relatively high density turbidite sandstones containing abundant top and middle absent Bouma sequences. Many of the thinner bedded sandstones and siltstones are finer grained and are much less laterally continuous; some of this variability may result from original depositional variability, though most appears to be due to post-depositional processes. Soft sediment deformation is particularly common in this facies and may result from liquefaction and/or shear by relatively dense flows. Sediment failure may have been triggered by shock waves and shear in front of and beneath a turbidity current, slope failure (though no unambiguous slumping was found) or possibly due to shock waves produced by seismic activity. Palaeocurrent indicators are rare in the Bronllwyd Grit Formation, but those which have been found suggest that turbidite flow was parallel to the axis of the basin and was probably from the northeast (Crimes 1970a; Wood 1974).

6.4 : Conclusions.

The Lower Cambrian succession of Arfon in general shows a vertical trend from volcanics to alluvial fan and braided fluvial deposits, above which there is a transgressive sequence of shallow marine deposits and basinal turbidites.

Deposition began with the rapid eruption of subaerial ashflow tuffs from ponded pyroclastic flows (Padarn Tuff Formation). Such a large-scale eruption probably emptied the magma chamber from which it was derived and resulted in caldera collapse. Sediment was eroded off the upfaulted blocks into the subsiding grabens as alluvial fan conglomerates and braided fluvial sandstones (lower Minffordd Formation and the lower part of the Fachwen Formation). A few ash-flow and airfall tuffs indicate that vulcanicity continued above the Padarn Tuff Formation, though primary volcanic deposits are less common upwards relative to redeposited volcaniclastics. The Fachwen and Minffordd Formations in general fine upwards and the Fachwen Formation shows a transition from subaerial to shallow marine deposition up into dominantly fine grained basinal deposits in the overlying Llanberis Slates Formation. Isolated turbidite sandstones occur within the Llanberis Slates Formation but a sand-rich turbidite facies appears suddenly in the Bronllwyd Grit Formation. The Bronllwyd Grit Formation contains structures e.g. multiple grading which indicate deposition from high density turbidity currents.

Facies in the Fachwen Formation are laterally discontinuous. As described by Reedman *et al.* (1984) the Padarn Tuff Formation in the Bigil Syncline is overlain by conglomerates and a thick sequence of Fachwen Formation whereas about 1km to the east at Gallt-y-Foel the Padarn Tuff Formation is overlain only by a thin sequence of sandstones and siltstones. Lateral variations in facies indicate a complex pattern of sedimentation, probably with fault activity controlling deposition. Comparison of sequences between the Padarn and Bangor Ridges and even along strike within the Padarn inlier shows considerable variation in thickness and type of deposits. An unconformity occurs between the Minffordd and Bangor Formations in the Bangor inlier but no such feature is seen in the Padarn inlier. These lateral variations indicate that there was active fault movement on the Aber-Dinlle Fault at this time with upthrow to the northwest. A thicker sequence of conglomerates and sandstones at Nantlle compared to Llanberis (both in the Padarn inlier) suggests maximum subsidence at this time in the southwest.

The strata appear to become more laterally continuous in the upper part of the sequence, though in the Bangor inlier much of the upper Lower and Middle Cambrian succession has been removed by pre-Arenig erosion. The increased lateral continuity in the younger strata indicates that irregular topography had by this time largely been filled and the Arfon Basin underwent more general subsidence, resulting in a deepening of the basin. Thus while the Llanberis Slates Formation was being deposited, the Arfon Basin was a deep, turbidite basin.

Fault activity still controlled sedimentation, however (Wood 1974; Webb 1983). Sandy turbidites in the Llanberis Slates Formation and Bronllwyd Grit Formation show thickness variations which Webb (1983) interpreted being fault controlled, though overall sand packets have a tabular geometry and are laterally continuous.

CHAPTER SEVEN

DETRITAL PETROGRAPHY

CHAPTER SEVEN : DETRITAL PETROGRAPHY.

Introduction.

The petrography of parts of the Lower and Middle Cambrian of North Wales has already been described in detail in the literature (e.g. Matley & Wilson 1946; Okada 1966, 1967; Binstock 1977). The aim of this petrographic study was to use the criteria of Dickinson & Suczek (1979) in order to define the plate tectonic setting of the Welsh Basin at this time and to attempt to differentiate between sediment sources during the progressive infill of the basin.

Method.

The detrital composition of coarse grained rocks from the Dolwen Formation (4), Rhinog Formation (36), Hafotty Formation (2), Barmouth Formation (15), Gamlan Formation (14), Cefn Coch Grit (2), Hell's Mouth Grits (17), Mulfran Beds (1), Cilan Grits (12) and Bronllwyd Grit Formation (11) were examined in thin section and their compositions quantitatively determined (500 points per thin section). Sandstones and some conglomerates were sampled in order to compare similar grain sizes and identify the type(s) of source(s) the different formations were derived from. The sample is biased towards the coarser grain sizes which predominantly occur near the bases of T_a turbidite beds.

Thin sections were stained for potash feldspar using the method of Wilson & Sedeora (1979). Individual feldspar grains and feldspars within lithic fragments were analysed in five polished thin sections (two from the Rhinog Formation and one each from the Barmouth Formation, Hell's Mouth Grits and the Bronllwyd Grit Formation) using an energy dispersive microprobe system housed at the University of Manchester.

The majority of Lower and Middle Cambrian sandstones in North Wales are lithic and feldspathic greywackes (Fig 7.1, filled circles), using the scheme of Pettijohn *et al.*

(1973). However some scour fills, sets of cross-bedding and, as observed by Okada (1966), some coarser bases to T_a beds contain non-greywacke sandstones (sublitharenites and lithic arenites, Fig 7.1, open circles).

The following grain types are present:

1) <u>Quartz.</u>

Monocrystalline quartz is usually the most common framework grain type in the samples examined. It may have uniform extinction, though undulose extinction is more common. Polycrystalline quartz is also common; there are several types (c.f. Folk 1974) including those with straight boundaries between crystals (possibly igneous), rounded (sedimentary) and sutured, often containing elongate crystals (metamorphic). Some grains are observed to contain small sub-grains around their outer margin. These textures suggest sub-grain formation during the metamorphism of the Cambrian succession. Thus the amount of polycrystalline quartz may be overestimated with respect to the original detrital composition. In general there are few consistent differences in the proportion of different types of quartz between formations.

Most grains are sub-rounded to sub-angular though they range from angular to well rounded. Many grains have jagged, irregular edges and some have re-entrant outlines, which are most common in the monocrystalline grains.

2) <u>Feldspar.</u>

Feldspar is common in many of the Lower and Middle Cambrian sandstones in North Wales, though it is much less common than quartz. Lamellar and untwinned feldspars are most common though some simple twinned and perthitic grains are also present. The simple twinned, prismatic feldspars are similar to crystals within many of the volcanic lithic fragments. Feldspars range from highly altered to fresh even within the same thin section. Many patches of chlorite appear to have been concentrated around feldspar grains.

The relative proportions of total feldspar are similar between formations, though the Bronllwyd Grit Formation is relatively rich in lamellar twinned feldspars compared to the Rhinog Formation.

Okada (1967) and Binstock (1977) suggested that the sandstones from the Harlech Grits Group contained a mixture of plagioclase and potash feldspar grains. 95 out of 97 points analysed by microprobe were however found to be albitic in composition. The mean feldspar composition is 97.8 mole% albite and several feldspars contain 100 mole% albite. There are few differences in feldspar composition between formations, the highest mean occurs in the Hell's Mouth Grits (99.5 mole% albite) and the lowest in the Barmouth Formation (97.3 mole% albite) and the ranges overlap considerably.

Different types of feldspar were analysed- lamellar, simple and untwinned. However little difference in albite composition was detected between types or within grains.

A relatively small number of oligoclase grains are also present.

Detrital feldspars are commonly derived from either igneous or metamorphic rocks (Folk 1974). Unaltered igneous rocks containing feldspars which are predominantly albite are extremely uncommon though albite is common in low grade metamorphic rocks (Kastner & Siever 1979). Therefore the abundance of albite in the Lower and Middle Cambrian sandstones can be accounted for in several ways, by:

i) Segregation of albite from other feldspars by weathering processes and abrasion during transport. Calcic and some potash feldspars are generally less stable at near surface temperatures and pressures and therefore may be more easily attacked chemically and physically than albitic feldspars (Kastner & Siever 1979). Albitic feldspar may survive to be deposited in the turbidite basin, though it seems unlikely that such complete segregation would occur.

ii) Supply of sediment from an albitic source, i.e. one that has been albitised by metamorphism or hydrothermal activity (?Anglesey).

iii) *In situ* albitisation. The Lower and Middle Cambrian sediments may originally have had variable feldspar compositions. Dissolution of plagioclase and replacement by albite is particularly common during deep burial diagenesis of feldspathic sandstones and greywacke sequences (Kastner & Siever 1979), e.g. beneath the Gulf Coast, U.S.A. (Gold 1987, Land & Milliken 1981), and similar processes may have affected the Lower and Middle Cambrian sandstones of North Wales. Albite is very common in greywackes, especially those that are Palaeozoic or earlier in age and have been subjected to high grade diagenesis/low grade regional metamorphism (Kastner & Siever 1979). Since Cambrian rocks in North Wales have suffered low grade regional metamorphism (Bevins & Rowbotham 1983) this may have caused the widespread albitisation of feldspars.

3) Lithic Fragments.

Lithic fragments are generally more common in the coarser grained deposits. A variety of types of lithic fragments occur (e.g. Matley & Wilson 1946; Greenly 1919):

a) Volcanic. Volcanic clasts are common within the Arfon Group and are mainly locally derived. Elsewhere they are less common, though some coarser beds in the Rhinog Formation (often associated with sets of cross-bedding) are comparatively rich in volcanic clasts (Plate 7/I and II). The volcanic clasts are quartz-poor and are probably intermediate or basic in composition. They are usually composed of laths of feldspar; some types show trachytic texture and/or simple twinned feldspar phenocrysts. Feldspars are dominantly albitic; 35 out of 37 points were albite and had similar compositions to the isolated feldspar grains, so both types have probably been albitised. There is considerable variation in crystal size and some of the coarser varieties may be derived from hypabyssal intrusions. Some tuffaceous clasts also occur and Crimes (1970a) reports the presence of flow-banded rhyolite clasts in the Rhinog Formation. In other formations they are usually much less common, especially the Barmouth and Gamlan Formations.

b) Plutonic. Most plutonic lithic clasts consist of microgranite, some of which show micrographic (Plate 7/III) texture. Plutonic clasts occur in most formations but are particularly common in the Barmouth, Gamlan and Bronllwyd Grit Formations and parts of the Cilan Grits. Many beds in the Rhinog Formation are relatively poor in plutonic lithic clasts.

c) Sedimentary/Metamorphic. Sedimentary quartzite clasts are common as well as siltstone and mudstone fragments (?intraclasts). Metamorphic clasts are sometimes common, usually of metaquartzite with elongate crystals. Grains containing quartz with undulose extinction and suturing at crystal boundaries indicate that they have been highly strained (Plate 7/IV). Occasional grains of cataclasite also occur. Quartz-rich metamorphic grains are particularly common in some beds in the Cilan Grits and Bronllwyd Grit Formation but also occur in other formations.

4) Micas.

Muscovite mica and chlorite are relatively common.

5) Heavy Minerals.

Detrital grains of zircon are occasionally observed in samples. Opaque heavy minerals are restricted to magnetite.

Matrix.

The matrix is largely composed of chlorite and sericite with occasional epidote and probably formed as a result of burial diagenesis/low grade metamorphism. The amount of matrix varies, though most sandstones are texturally immature and have about 20-25% matrix with a maximum of about 50%. Some samples however contain less than 15% matrix; these were usually collected from coarser deposits. The origin of the matrix in greywackes has been a matter of controversy (Cummins 1962) but it seems likely that the matrix was formed by metamorphic alteration of feldspars, lithics (which were possibly of volcanic origin, c.f. Brenchley 1969) and original clay minerals. However there is little evidence to indicate the origin of much of the matrix in these rocks and the point count data probably underestimate the original proportion of detrital labile grains.

Provenance.

Several plots were made in order to compare the different formations with the plots of Dickinson & Suczek (1979) and Folk (1974) (Figs 7.1 to 7.5). It is difficult to differentiate between formations though the Rhinog Formation contains some sandstones which are particularly rich in volcanic clasts (Fig 7.4).

In general parts of the Rhinog Formation and to a lesser extent the Hell's Mouth Grits are richer in volcanic lithics while the Barmouth Formation, Gamlan Formation, Bronllwyd Grit Formation and parts of the Cilan Grits tend to be comparatively rich in plutonic clasts. Parts of the Cilan Grits and Bronllwyd Grit Formation commonly contain metamorphic clasts. Gibbons (1983a) suggested that metamorphic clasts in the Cilan Grits indicated a Monian source.

Figs 7.1 to 7.3 are based on Dickinson & Suczek's (1979) plate tectonic provenance types which are summarised in Fig 7.6. These plots indicate that the Lower and Middle Cambrian sediments of North Wales were derived mainly from a recycled orogen. However Fig 7.3, when compared with Fig 7.6c suggests derivation either from a subduction complex or more likely a mixed supply of collision orogen and arc orogen. This is probably because much of the sedimentary/metamorphic material is biased towards detrital polycrystalline quartz grains.

Okada (1967) found a variety of heavy minerals in the Harlech Grits including euhedral and rounded zircon, tourmaline, garnet and monazite. This indicates a mixed acid and basic igneous with a minor metamorphic source (Binstock 1977). The presence of euhedral zircons indicate that some of the sediment was first cycle.

In summary, the petrography indicates that the Lower and Middle Cambrian rocks of North Wales were probably supplied from locally derived volcanics in their lower part (see chapter 6) and from a recycled orogen in their upper part. The source of the sediment was a mixed metamorphic/plutonic source to the north (?Anglesey) for the Bronllwyd Grit Formation, Cilan Grits and parts of the Rhinog Formation and Hell's Mouth Grits. The Barmouth and Gamlan Formations were derived from a southerly source, richer in plutonic lithic clasts. The Rhinog Formation may have been derived from both quartz-rich (to the north) and basic/intermediate volcanic sources (to the east/northeast). The high proportion of relatively angular quartz grains (e.g. Plate 7/V) and the presence of compositionally immature grains (e.g. volcanic fragments) indicate that they are first cycle and the distance between the source and the basin of deposition was small.Therefore any shallow marine shelf between the source and the basin where turbidites were deposited was relatively narrow or there was a way that the shelf could be bypassed.

The sources of the Lower and Middle Cambrian of North Wales were essentially similar reworked orogen with only some variation in the proportion of reworked volcanics. The nature of the source and the occurrence of shallow marine deposits on a stable shelf to the south and east of North Wales suggest that it is unlikely that the Lower and Middle Cambrian were derived from the Midland Platform.

FIGS 7.1 to 7.6 Petrographic triangular diagrams for the Rhinog, Barmouth and Gamlan Formations, Hell's Mouth Grits, Cilan Grits and Bronllwyd Grit Formation.

KEY

Q : Total Quartz

Qm : Monocrystalline Quartz

Qp : Polycrystalline Quartz

F : Feldspar

L : Total Lithic Fragments

Lv : Volcanic Lithic Fragments

Lp : Plutonic Lithic Fragments

Ls : Sedimentary and Metamorphic Lithic Fragments

⊕ Arenites

● Greywackes

FIG 7.1 a

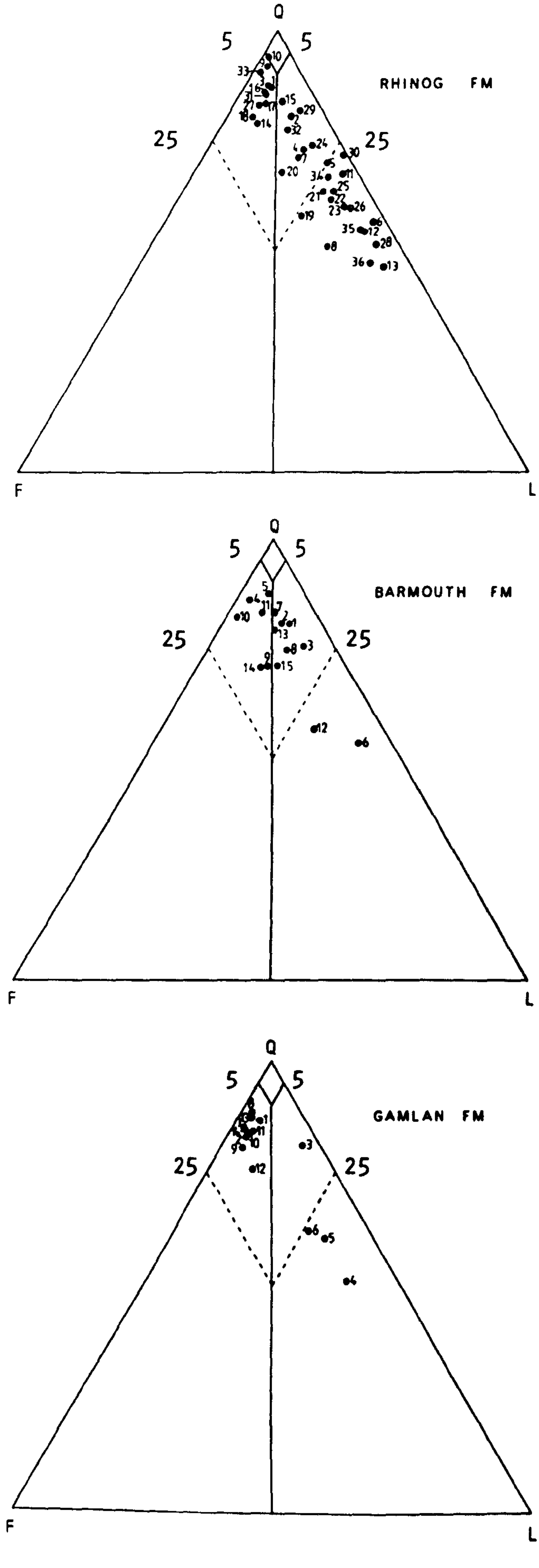

Q
5
5
25
25
RHINOG FM
F
L
Q
5
5
25
25
BARMOUTH FM
F
L
Q
5
5
25
25
GAMLAN FM
F
L

FIG 7.1b

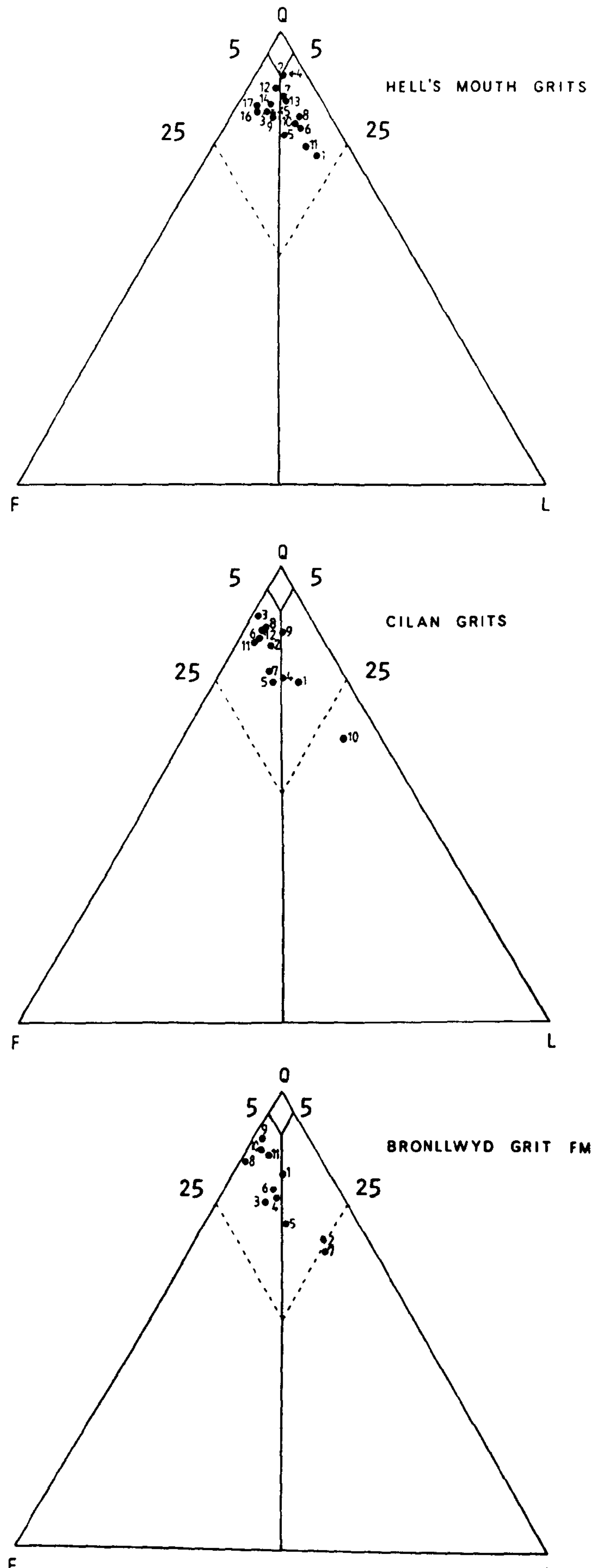

Q
5
5
25
25
HELL'S MOUTH GRITS
F
L
Q
5
5
25
25
CILAN GRITS
F
L
Q
5
5
25
25
BRONLLWYD GRIT FM
F
L

FIG 7.2a

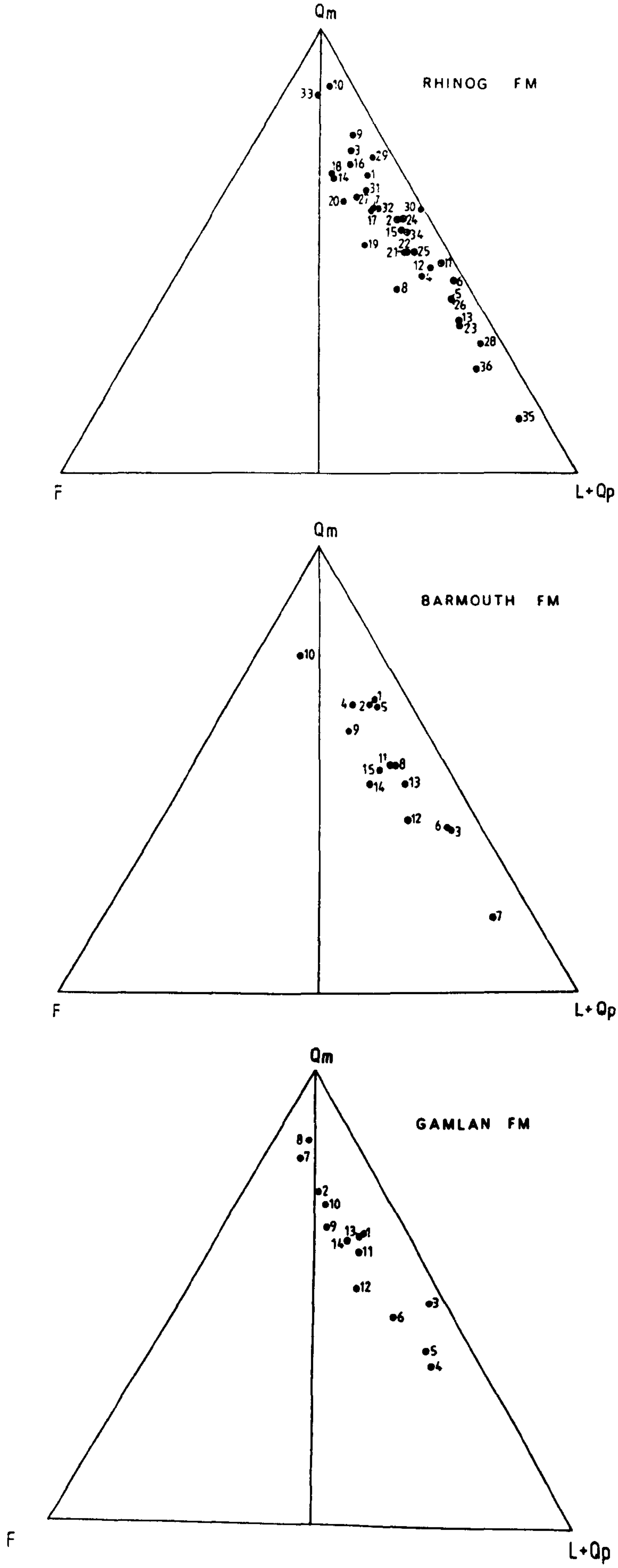


FIG 7.2b

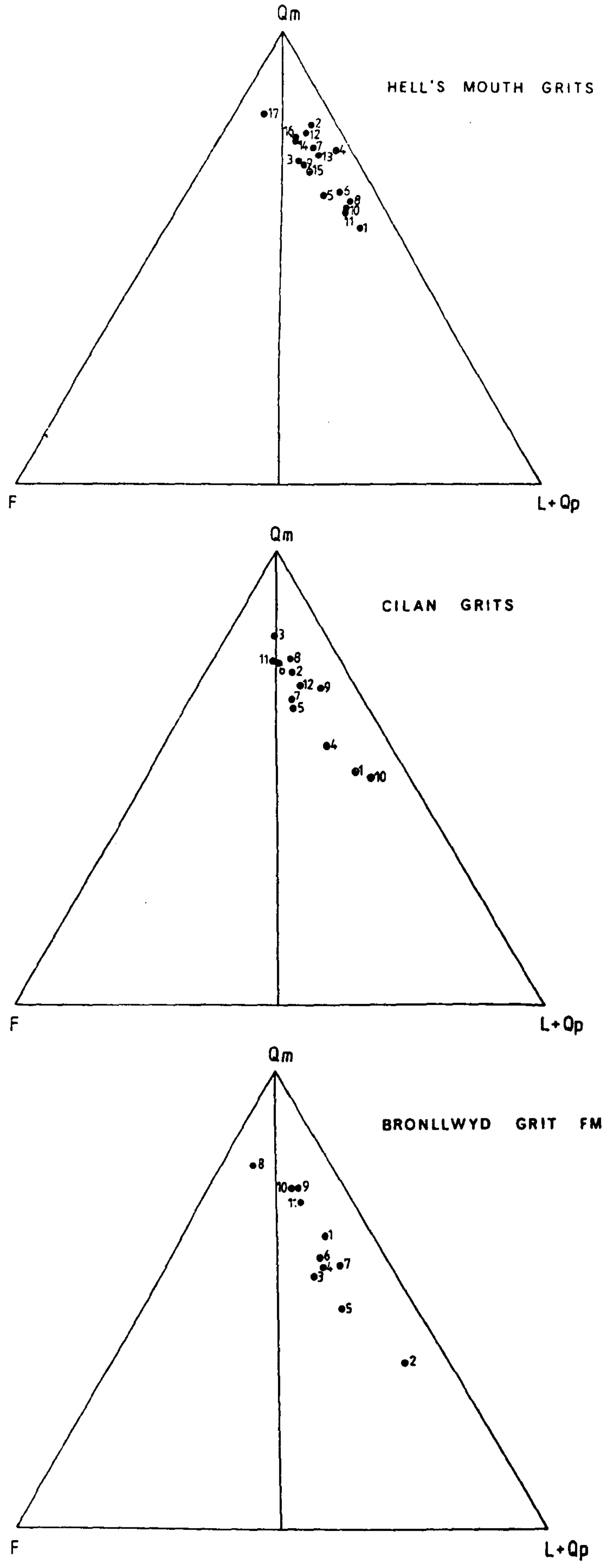

Qm
HELL'S MOUTH GRITS
F
L+Qp
Qm
CILAN GRITS
F
L+Qp
Qm
BRONLLWYD GRIT FM
F
L+Qp

FIG 7.3a

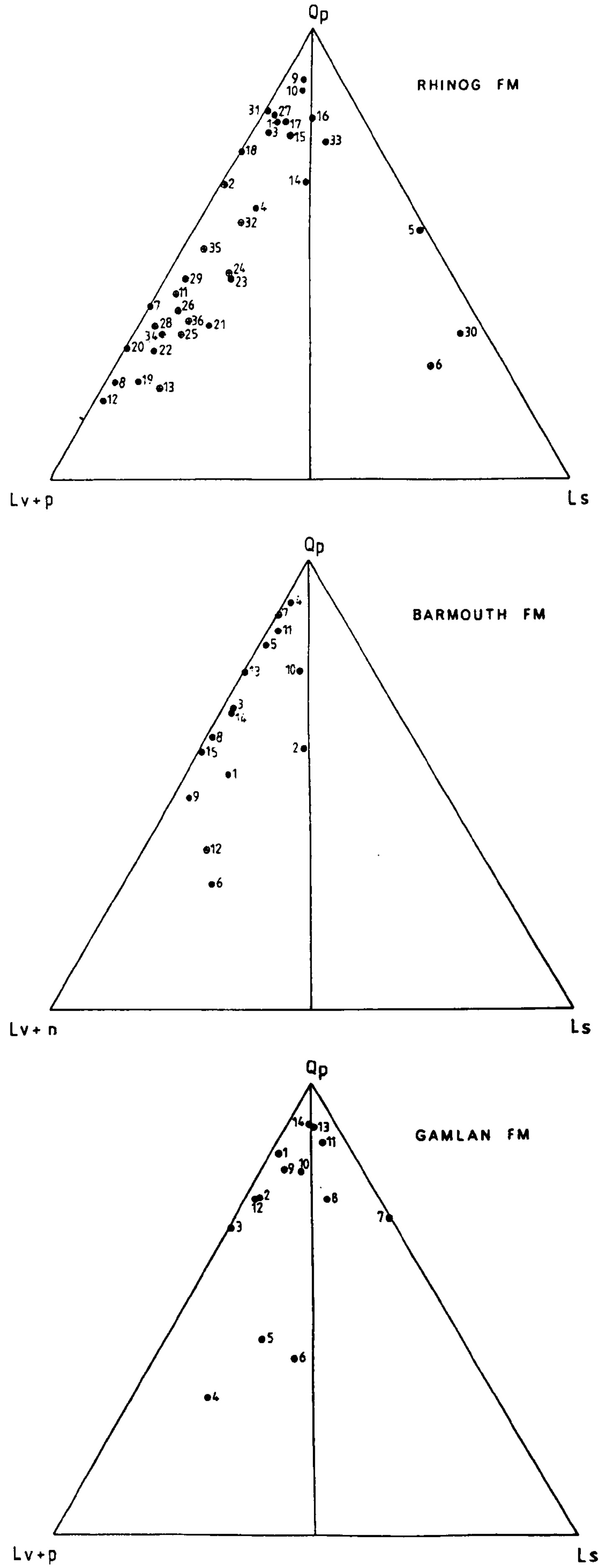


FIG 7.3b

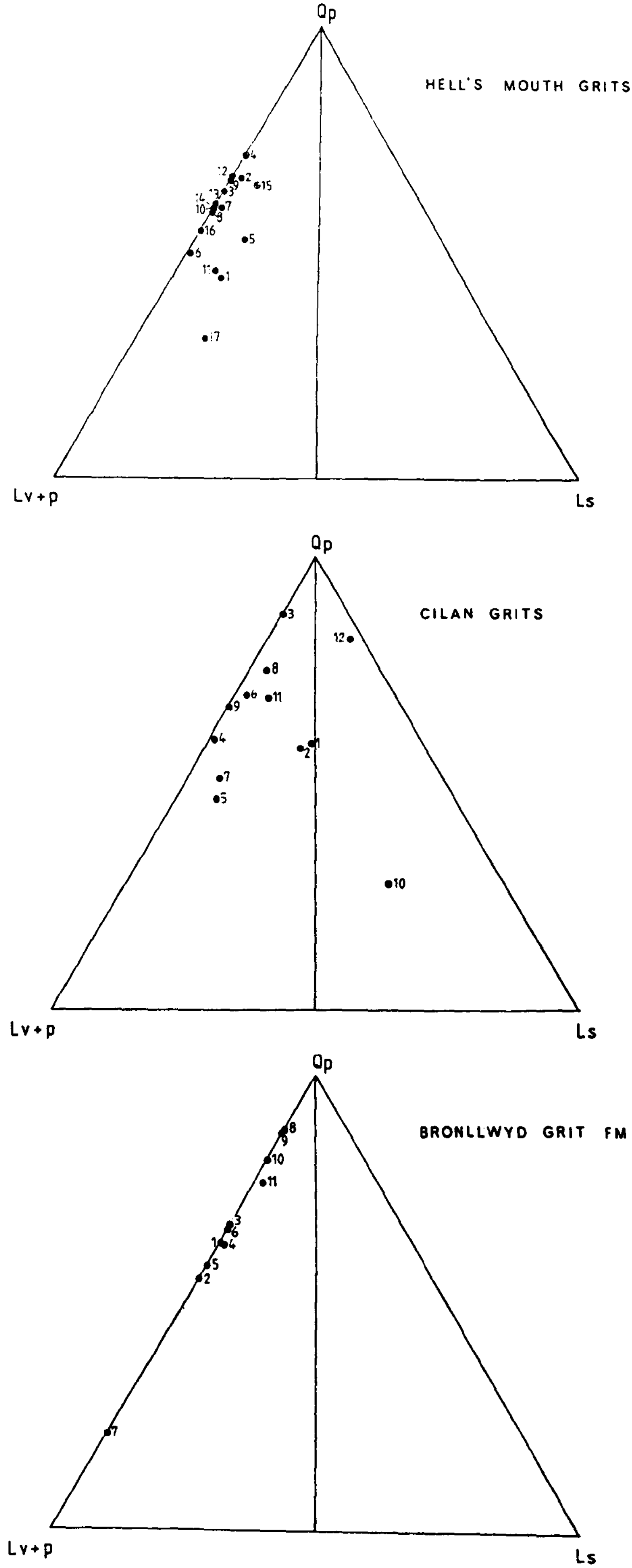

Qp
HELL'S MOUTH GRITS
Lv+p
Ls
Qp
CILAN GRITS
Lv+p
Ls
Qp
BRONLLWYD GRIT FM
Lv+p
Ls

FIG 7.4a

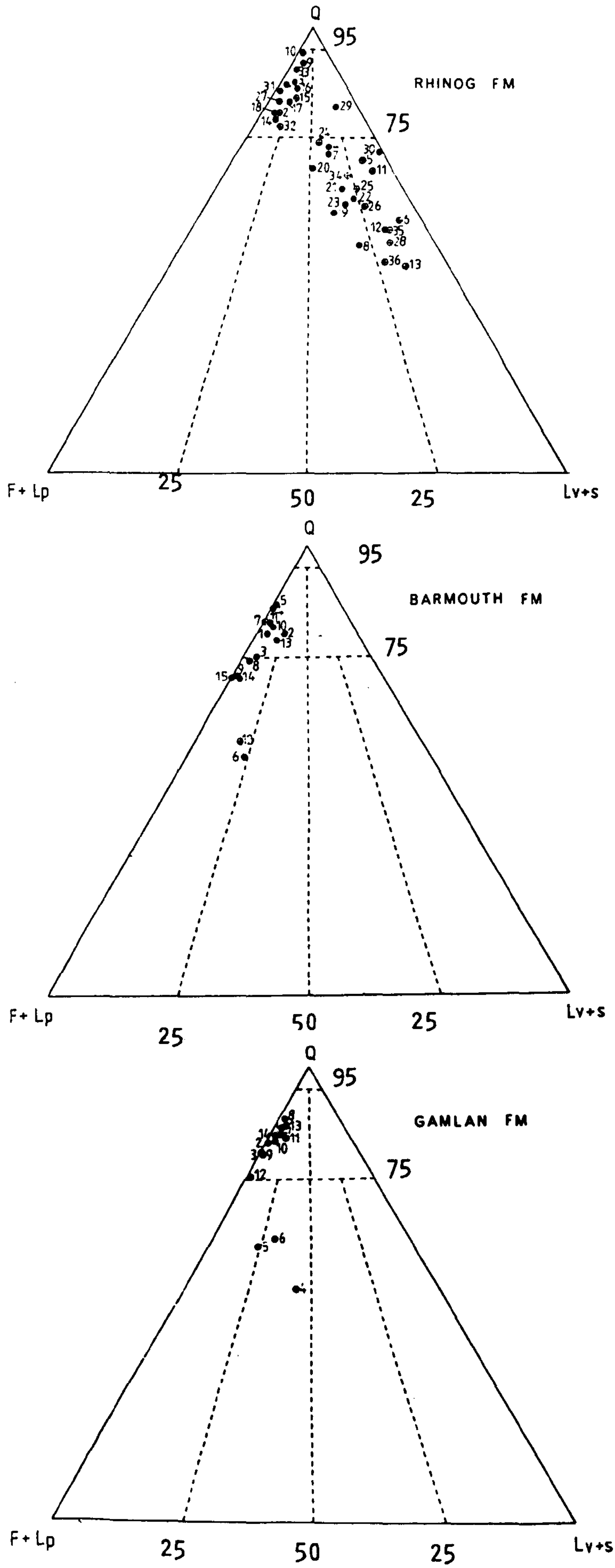

Q
95
RHINOG FM
75
F+ Lp
25
50
25
Lv+s
Q
95
BARMOUTH FM
75
F+ Lp
25
50
25
Lv+s
Q
95
GAMLAN FM
75
F+ Lp
25
50
25
Lv+s

FIG 7.4b

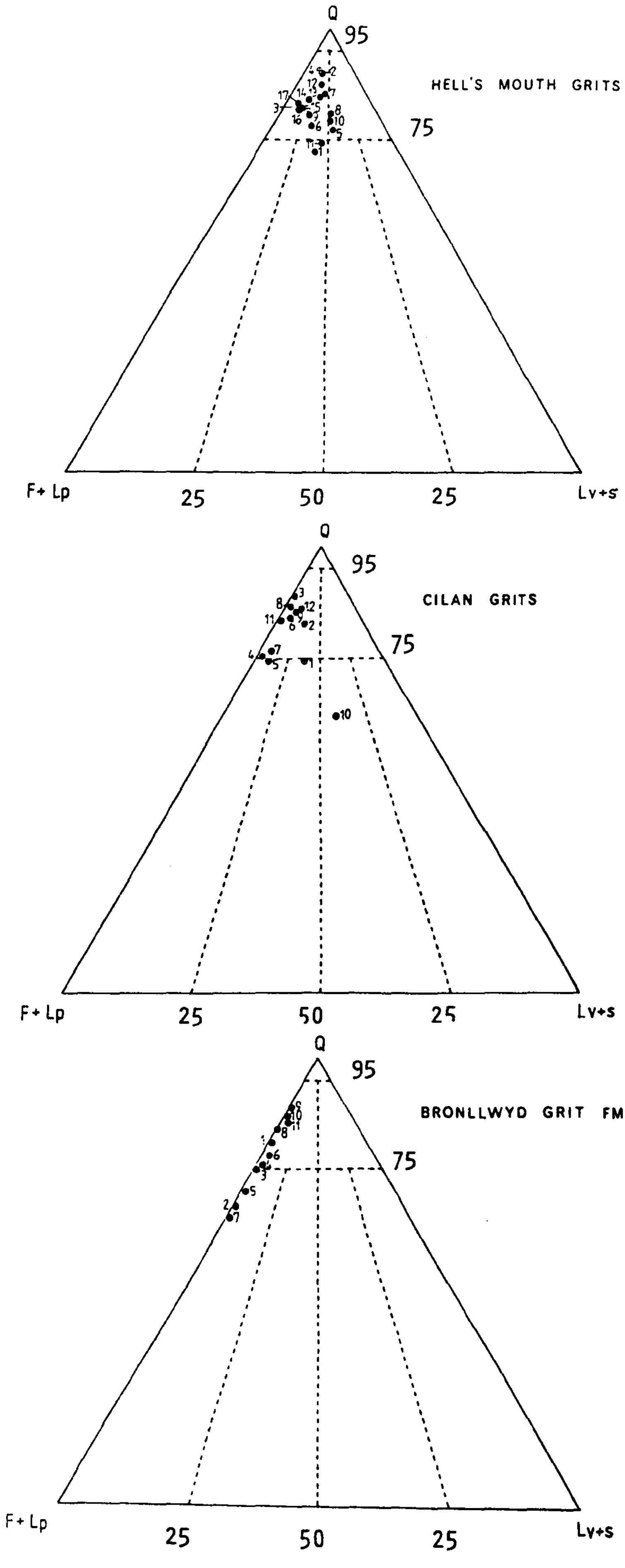

Q
95
HELL'S MOUTH GRITS
75
F+ Lp
25
50
25
Lv+s
Q
95
CILAN GRITS
75
F+ Lp
25
50
25
Lv+s
Q
95
BRONLLWYD GRIT FM
75
F+ Lp
25
50
25
Lv+s

FIG 7.5a

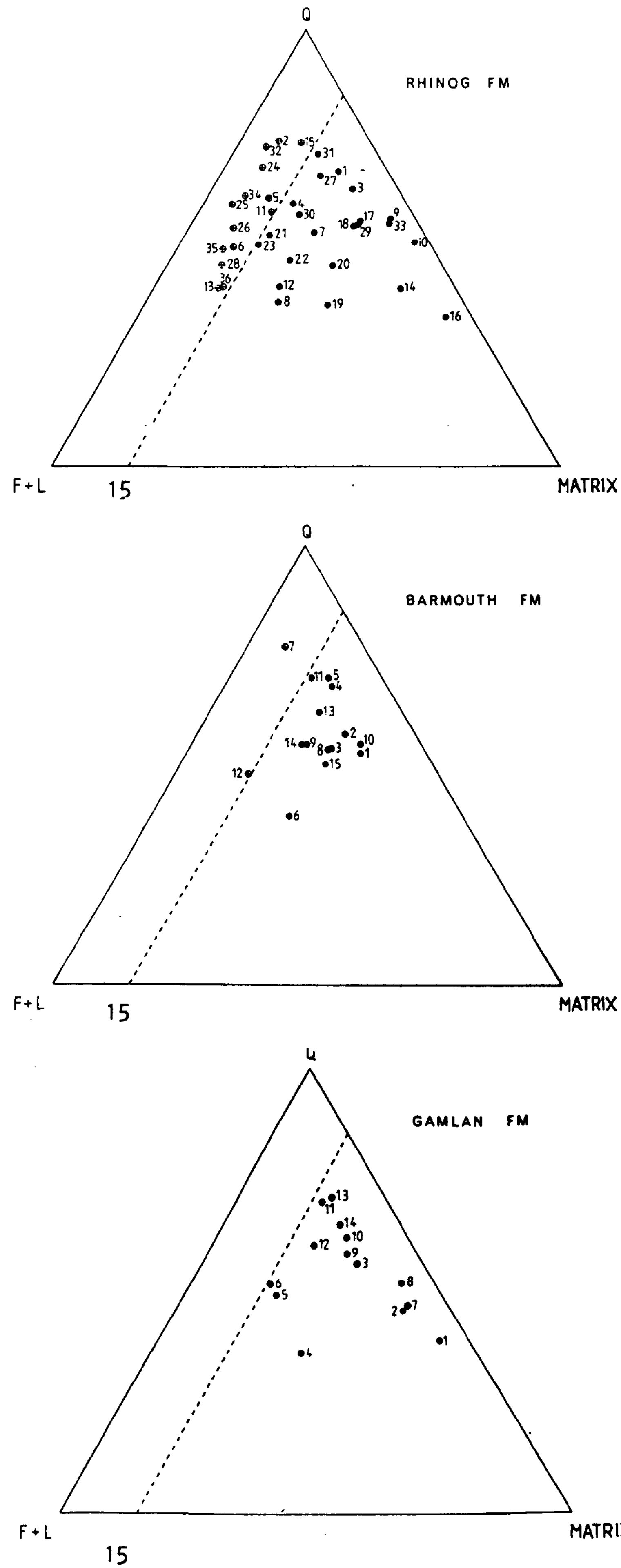

Q
RHINOG FM
F+L
15
MATRIX
Q
BARMOUTH FM
F+L
15
MATRIX
Q
GAMLAN FM
F+L
15
MATRIX

FIG 7.5b

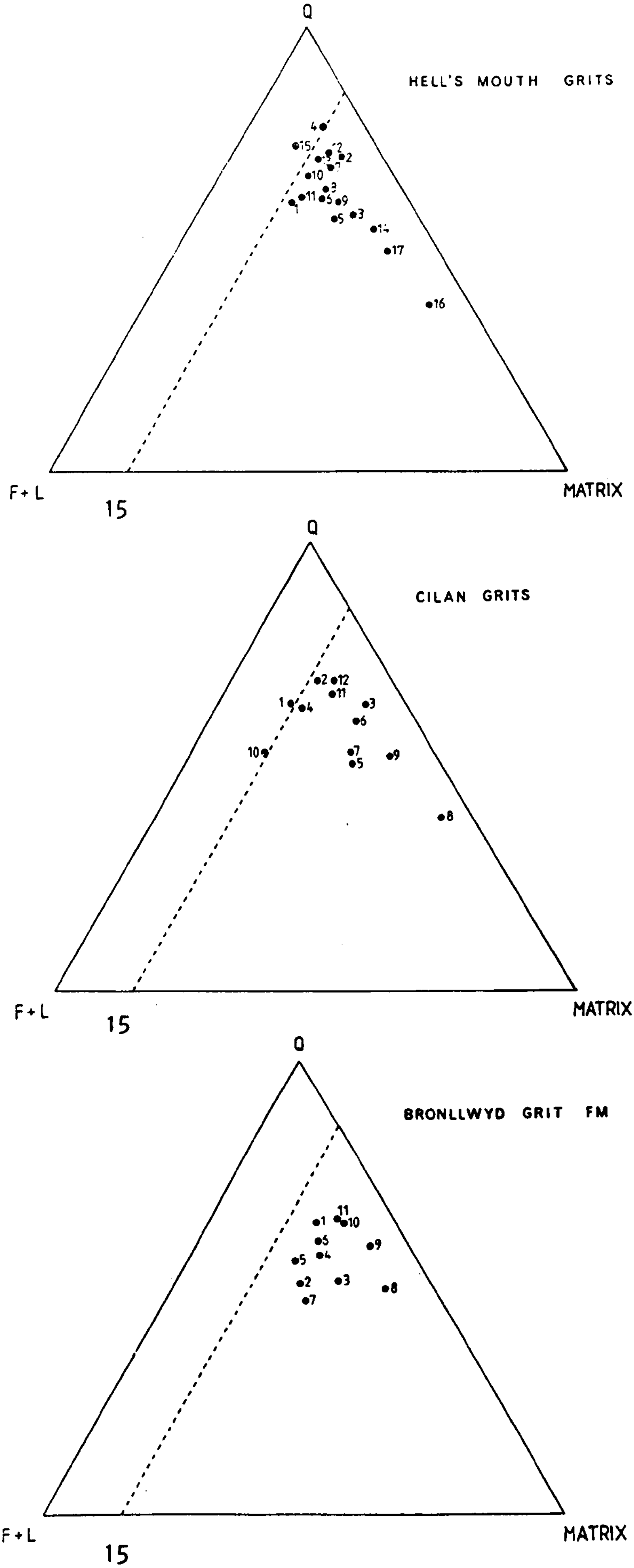
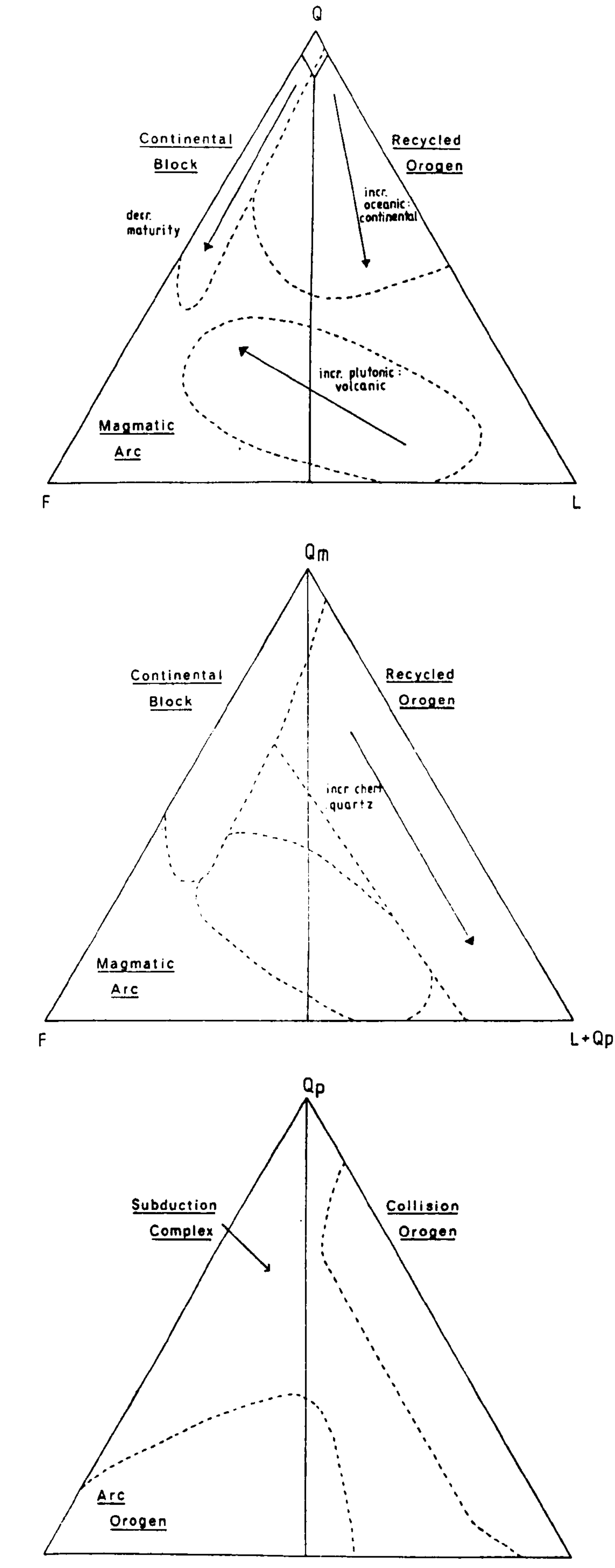

Q
HELL'S MOUTH GRITS
F+L
15
MATRIX
Q
CILAN GRITS
F+L
15
MATRIX
Q
BRONLLWYD GRIT FM
F+L
15
MATRIX

FIG 7.6



after Dickinson and Suczek (1979)

PLATE 7/I : Thin section (cross polars) of volcanic clasts in the Rhinog Formation.

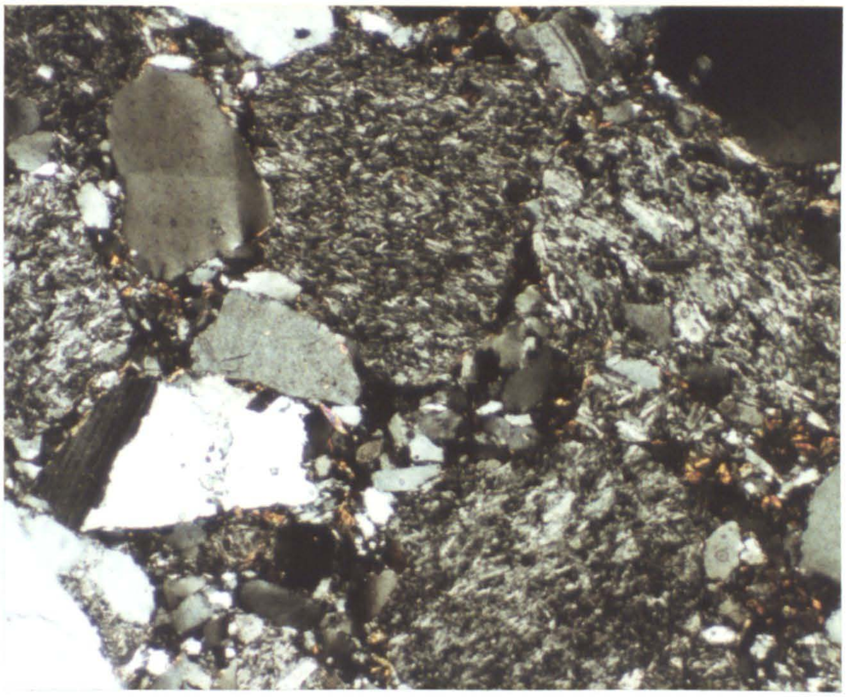


PLATE 7/II : Thin section (cross polars) of a volcanic clast showing trachytic texture, Rhinog Formation.

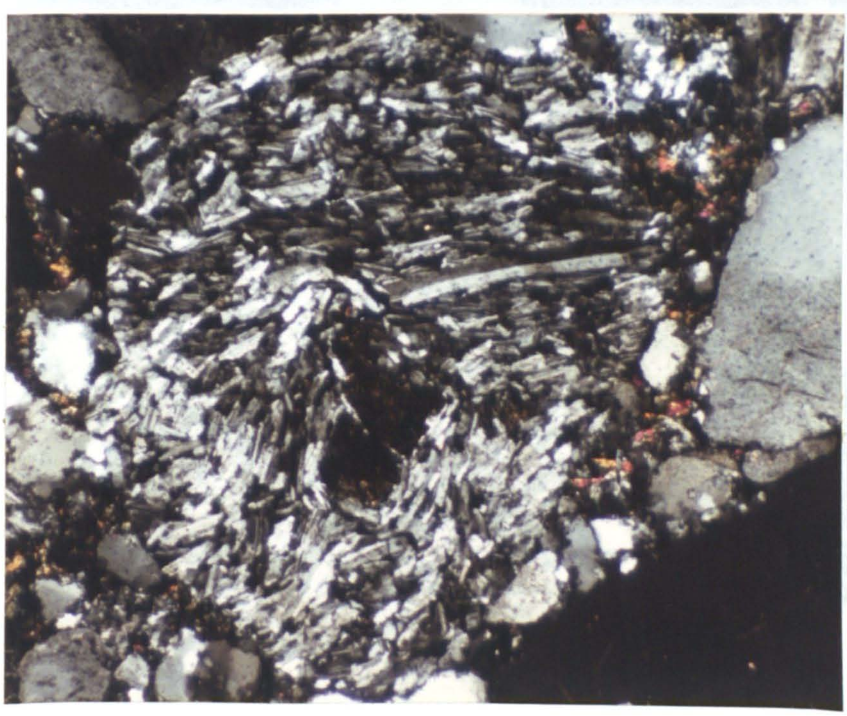


PLATE 7/III : Thin section (cross polars) of a clast showing micrographic texture, Rhinog Formation.

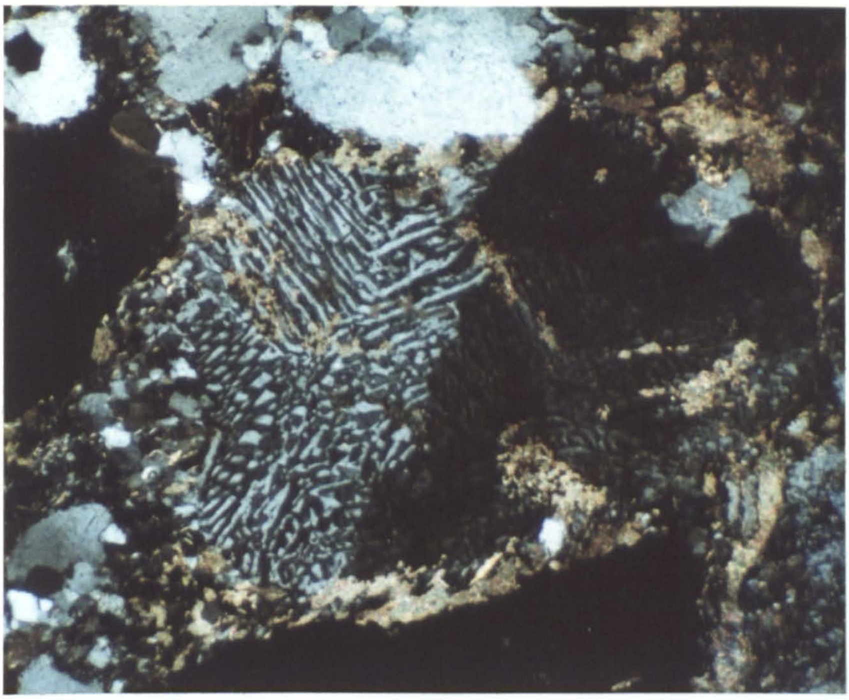


PLATE 7/IV : Metamorphic Quartzite grain, Cilan Grits.

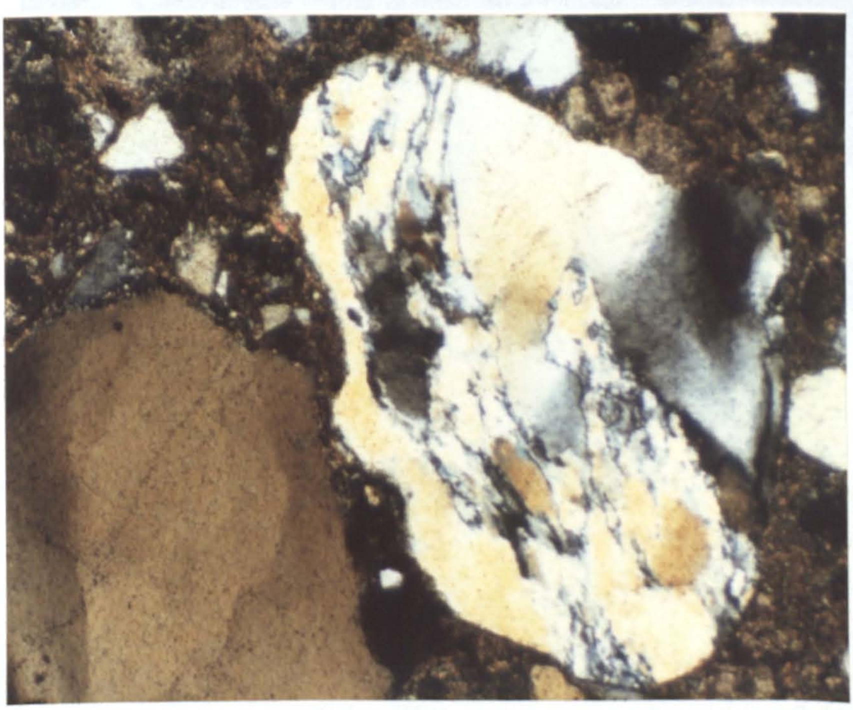
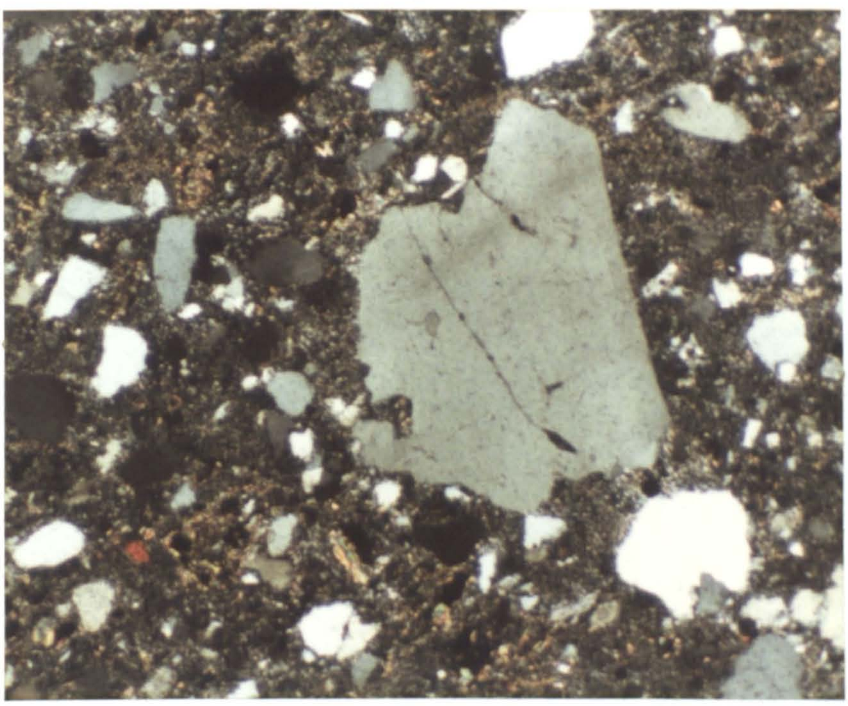


PLATE 7/V : Angular grains in greywacke sandstone, Rhinog Formation.



0 1mm

CHAPTER EIGHT

CONCLUSIONS

CHAPTER EIGHT : CONCLUSIONS.

The evolution of the Welsh Basin in the Lower and Middle Cambrian can be divided into the following stages (Figs 8.1, 8.2, 8.3):

1) Initial Volcanism.

After a period of oblique slip deformation (evidence from Anglesey and the Lleyn Peninsula, Gibbons 1983a,b) sedimentation began in North Wales with the eruption of volcanics. In Arfon a thick sequence of predominantly silicic, ponded ash-flow tuffs (Padarn Tuff Formation) were erupted within the NE-SW aligned, fault bounded Arfon Basin (Reedman *et al.* 1984). Most of the pyroclastic deposits were confined to the basin but some spilled out towards the west and are now represented by small outliers on Anglesey, e.g. the Bwlch Gwyn Tuff (Reedman *et al.* 1984).

In the Harlech Dome area the Bryn-teg Volcanic Formation was deposited- a sequence of basic and intermediate pyroclastic rocks (with island-arc, calc-alkaline andesite-dacite characteristics) and mass-flow volcaniclastic sediments derived from unstable slopes dominated by pyroclastic debris (Allen & Jackson 1978).

The differences in composition between volcanics in Arfon and the Harlech Dome indicate that the Padarn Tuff and Bryn-teg Volcanic Formations probably were not erupted synchronously. A conglomerate overlying the Bryn-teg Volcanic Formation contains acidic volcanic clasts similar to the Padarn Tuff and suggests that the Bryn-teg Volcanic Formation may be older (Allen & Jackson 1978). Differences in mode of deposition may also suggest that the two formations may have been deposited in separate basins- the Padarn Tuff Formation in the Arfon Basin (Reedman *et al.* 1984) and the Bryn-teg Volcanic Formation in the Merioneth Basin (new name). The boundaries of the Merioneth Basin are, however, poorly constrained.

2) Early Rifting Phase.

In Arfon eruption of the Padarn Tuff Formation was directly followed by differential subsidence and uplift, resulting in erosion and deposition of mainly acidic tuffaceous material. Later more basic volcanic material was deposited, probably sourced from Anglesey (Reedman *et al.* 1984). Lateral differences in sediment thicknesses and facies between sequences on the Bangor and Padarn Ridges on either side of the Aber-Dinlle Fault and even along strike on the Padarn Ridge between Llanberis and Nantlle indicate that deposition occurred within a system of fault confined sub-basins. The complicated tectonic framework led to complex facies distributions (Wood 1969; Reedman *et al.* 1984; present study). Conglomerate lenses are interpreted as alluvial fan deposits, probably derived from the erosion of local fault scarps, interbedded with, and laterally merging into braided fluvial deposits. In general the succession in the Arfon Basin fines upwards and the proportion of volcaniclastic sediment also decreases upwards. Sandstones higher in the succession (Fachwen Formation) and sponge spicules (Minffordd Formation, Reedman *et al.* 1984) indicate deposition in a shallow marine setting and reflect a general transgressive trend.

In the Merioneth Basin (Harlech Dome) sedimentation at this time was also represented by an upward transition from fluvial to shallow marine deposits (Dolwen Formation) but volcaniclastic deposits are very rare (Allen & Jackson 1985). Deposition in the upper part of the Dolwen Formation may have occurred within a deltaic environment (Allen & Jackson 1985 and present study) which may have been influenced by tidal currents.

3) General Deepening Phase.

If the Llanberis Slates Formation of Arfon can be correlated with the Llanbedr Formation of the Harlech Dome then generally fine grained deposition dominated in North Wales at this time. This probably reflects a general deepening in the Welsh Basin from shallow marine to one where basinal muds and subordinate turbidites accumulated. Occasional packets of sand-rich turbidites in the Llanberis Slates Formation are laterally continuous and may indicate deposition as wide sheets, though in detail faulting probably controlled the thickness of the deposits (Webb 1983).

4) First Phase of Sand-rich Turbidite Sedimentation.

In Arfon fine grained turbidite deposition (Llanberis Slates Formation) probably continued into the upper part of the Lower Cambrian.

In the Harlech Dome and St Tudwal's Peninsula the successions are different from the sequences in Arfon and probably indicate deposition in a separate basin or sub-basin (the Merioneth Basin). In the Harlech Dome the Rhinog Formation sand-rich turbidite system, derived mainly from the north, rapidly prograded over basinal shales (Llanbedr Formation). High density turbidites dominate. Although individual beds may have erosive bases and have a lensoid geometry, sand packets usually have a tabular geometry indicating possible deposition on weakly channelised lobes. Feldspathic and lithic greywackes are most common but some lithologies, often associated with scour fills and cross-bedding (frequently containing palaeocurrents indicating flow from the east), are particularly rich in volcanic clasts. The cross-bedding was produced by dilute, persistent, traction dominated flows capable of developing large bedforms. These flows may have originated from crevasse splay from a coarse grained, volcanic-rich, N-S elongated turbidite system to the east of the present outcrop or from a slope to the east (indicating closure of the Merioneth Basin to the east along possible N-S oriented faults).

The abundance of angular grains suggests that many of the grains were not greatly affected by abrasion prior to deposition. Thus in the Rhinog Formation sediment must have been rapidly eroded and transported rapidly over a narrow shelf. It seems unlikely that sediment bypassed a wider shelf since the occurrence of laterally equivalent sand-rich turbidites on St Tudwal's (Hell's Mouth Grits), derived from the northeast may indicate that turbidite deposition occurred on an apron rather than a fan fed by a single feeder channel or canyon.

The thick sandstone beds of the Hell's Mouth Grits of St Tudwal's are generally more laterally continuous than the Rhinog Formation. Conglomerate beds are present in the Rhinog Formation but absent in the Hell's Mouth Grits. The Hell's Mouth Grits may be the slightly more distal lobe deposits of the Rhinog Formation or, perhaps more likely, part of a separate turbidite system.

Above the top of the Rhinog Formation and Hell's Mouth Grits are manganiferous deposits which indicate a relatively abrupt "switching off" of the coarse clastic supply. This may have resulted from tectonic controls (c.f. Stow 1985) or a shifting of the locus of coarse grained sedimentation away from the outcrop area. The Mulfran Beds contain more sandy beds than their lateral equivalent the Hafotty Formation, suggesting that supply of sand continued, though to a lesser degree than before, in the St Tudwal's area.

5) Second Phase of Sand-rich Turbidite Sedimentation.

In Arfon the Bronllwyd Grit Formation indicates sand-rich turbidite deposition, possibly derived from the northeast (Crimes 1970a). Although overall the Bronllwyd Grit Formation has a tabular geometry, in detail contemporaneous faulting produced variations in thickness (Webb 1983).

In the Harlech Dome the Barmouth Formation was also deposited from high density turbidity currents, but derived from the south and was probably deposited, like the Rhinog Formation, in weakly channelised lobes on a sand-rich turbidite system. The source for this formation is problematical since Cambrian sediments in Shropshire and South Wales indicate stable shelf conditions. It is possible that post-Cambrian strike-slip (Woodcock 1984a,b) displaced the Welsh Basin with respect to its Cambrian sourcelands on its southern margin.

In St Tudwal's the Cilan Grits were derived from the northeast, contain evidence from high and low density turbidity currents and traction dominated flows flowing from the southeast.

The Barmouth Formation can be correlated with the Cilan Grits and indicate coarse grained turbidite deposition from the south and northeast respectively, indicating that the Welsh or Merioneth Basin was relatively narrow at this time. However the age of the Bronllwyd Grit Formation is not adequately constrained. Both the Bronllwyd Grit Formation and parts of the Cilan Grits contain metamorphic clasts which could have been derived from exposed Mona Complex on Anglesey, whereas the Barmouth Formation tends to be richer in plutonic clasts. Thus if the Bronllwyd Grit Formation is Middle Cambrian then the Bronllwyd Grit Formation and Cilan Grits may have been part of the same or similar turbidite systems. However on faunal grounds (see Chapter 3) the Bronllwyd Grit Formation may be upper Lower Cambrian and equivalent to the Rhinog Formation, so the two formations could have been part of the same turbidite system or more likely in separate basins or sub-basins. If the Bronllwyd Grit Formation is Upper Cambrian the Arfon was probably an area of non-deposition throughout much of the Middle Cambrian. Therefore the model postulated in Fig 8.1 (4,5) is only very tentative.

The Barmouth Formation is followed by the Gamlan Formation in the Harlech Dome, which indicates predominantly dilute, fine grained turbidite deposition derived from the south. On St Tudwal's the upper part of the Caered Mudstones indicate a possible upward transition into relatively quiet water shelf deposition.

In the Harlech Dome and St Tudwal's the sequences were then blanketed by dark, organic-rich shales and mudstones (Clogau Formation, Nant-pig Mudstones) which may indicate deepening as a result of a eustatic rise in sea level (Leggett 1980).

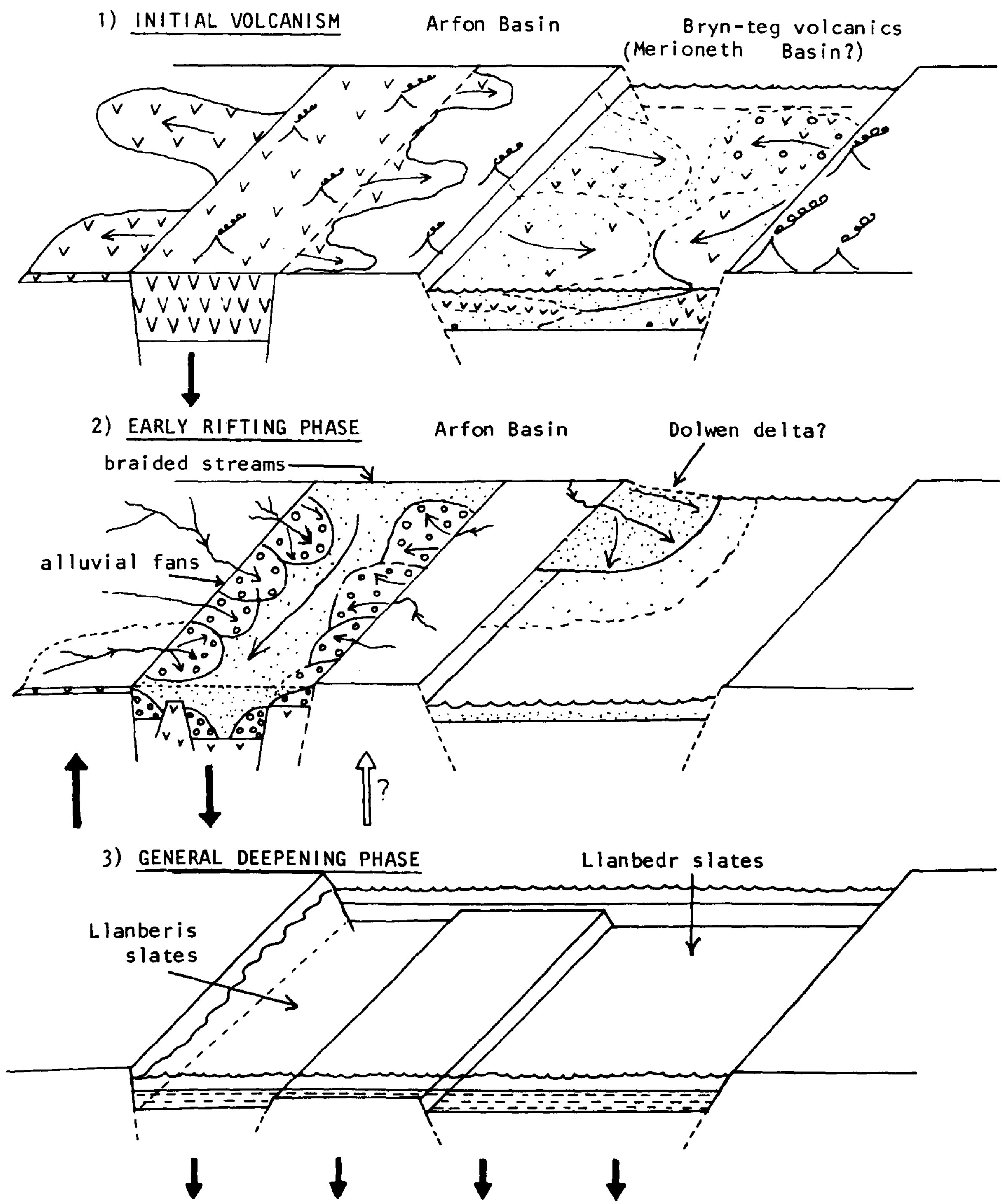
In summary the early part of the Lower Cambrian succession in North Wales indicates differential vertical fault movement within separate basins. Much of the initial subsidence in the Arfon Basin was probably volcanogenically controlled, resulting from caldera collapse after the extrusion of large volumes of pyroclastic material (c.f. Smith & Bailey 1968). Local subsidence probably continued, within laterally confined basins and sub-basins until in the Llanberis Slates and Llanbedr Formations more general subsidence occurred in North Wales to form a turbidite basin.

Sand-rich turbidite systems (e.g. Rhinog and Barmouth Formations) are particularly common in continental borderland basins such as in western California (Nilsen & Clarke 1975). However the occurrence of large volumes of volcanic material (Padarn Tuff Formation) and evidence which suggests that there was no strike slip movement on faults known to be active vertically during the Cambrian (e.g. the Dinorwic Fault, Reedman *et al.* 1984) suggests that a continental borderland setting is not appropriate for the Lower and Middle Cambrian rocks of North Wales. It is possible, however, that late Precambrian oblique strike slip deformation (Gibbons 1983a, b) produced an imprint on the structure of North Wales which was capable of being reactivated during later tectonic activity.

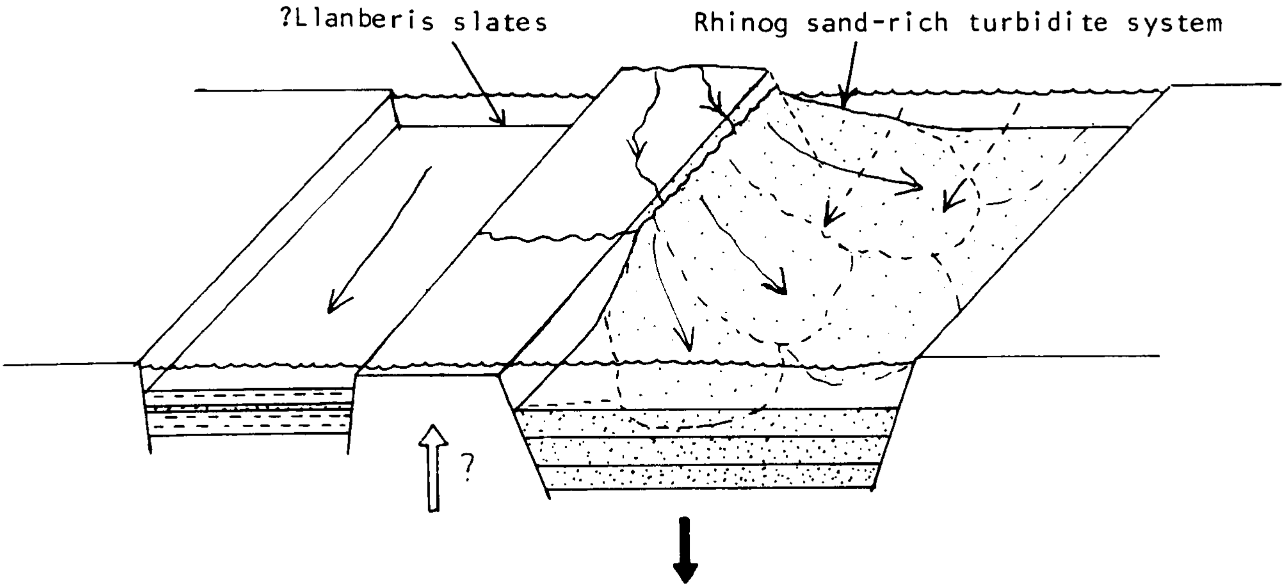
Instead the transition from locally confined subsidence to regional subsidence seems to fit better to the McKenzie (1978) model of crustal thinning due to stretching. McKenzie argued that under tension the lithosphere stretches and thins causing the asthenosphere to rise. At the surface this may result in volcanics and sedimentation in narrow grabens with rapid local subsidence (phases 1 and 2, Fig 8.1) and this is equivalent to the Rifting Phase. Rifting is followed by regional sagging (phases 3 to 5) caused by local cooling of the asthenosphere (Leeder 1982). It is possible that the Lower Cambrian crust of North Wales was under tension as a result of the break-up of the supercontinent referred to by Piper (1982) and Anderton (1980). However tension could have been subduction related (by correlation of the Bryn-teg Volcanic Formation with the volcanic island-arc deposits of the New Harbour Main and Conception Groups in the Avalon region of Newfoundland, Allen & Jackson 1978).

The pattern of sedimentation in North Wales is complicated by differential uplift and subsidence, which results in a switch of source in the Harlech Dome succession from northerly supplied sediments (Rhinog Formation) to southerly supplied sediments (Barmouth Formation). Subsidence probably decreased in the late Middle Cambrian and Upper Cambrian since these sediments show an overall regressive trend towards shallower water shelf deposits (Crimes 1970a).

FIG 8.1 Evolution of North Wales (tentative) in the Lower and Middle Cambrian.



4) FIRST PHASE OF SAND-RICH TURBIDITE SEDIMENTATION



5) SECOND PHASE OF SAND-RICH TURBIDITE SEDIMENTATION

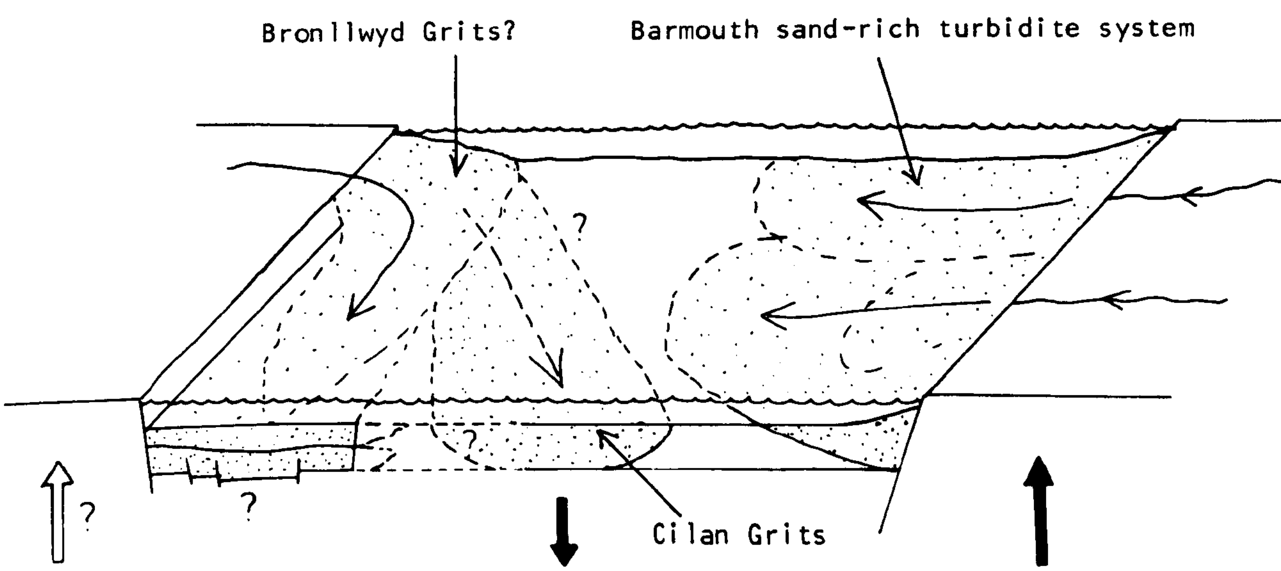


FIG 8.2 The Welsh Basin during the Lower and Middle Cambrian.

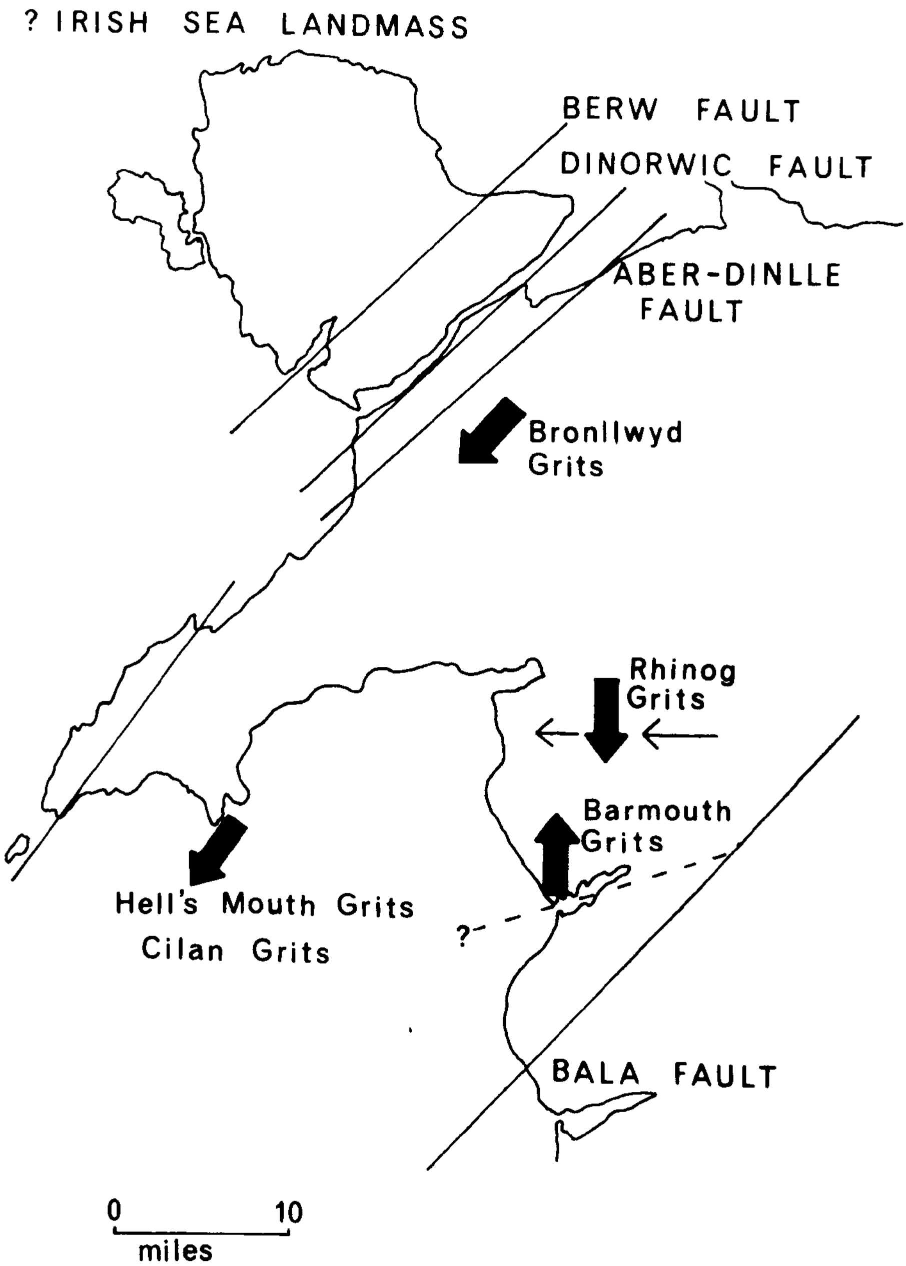
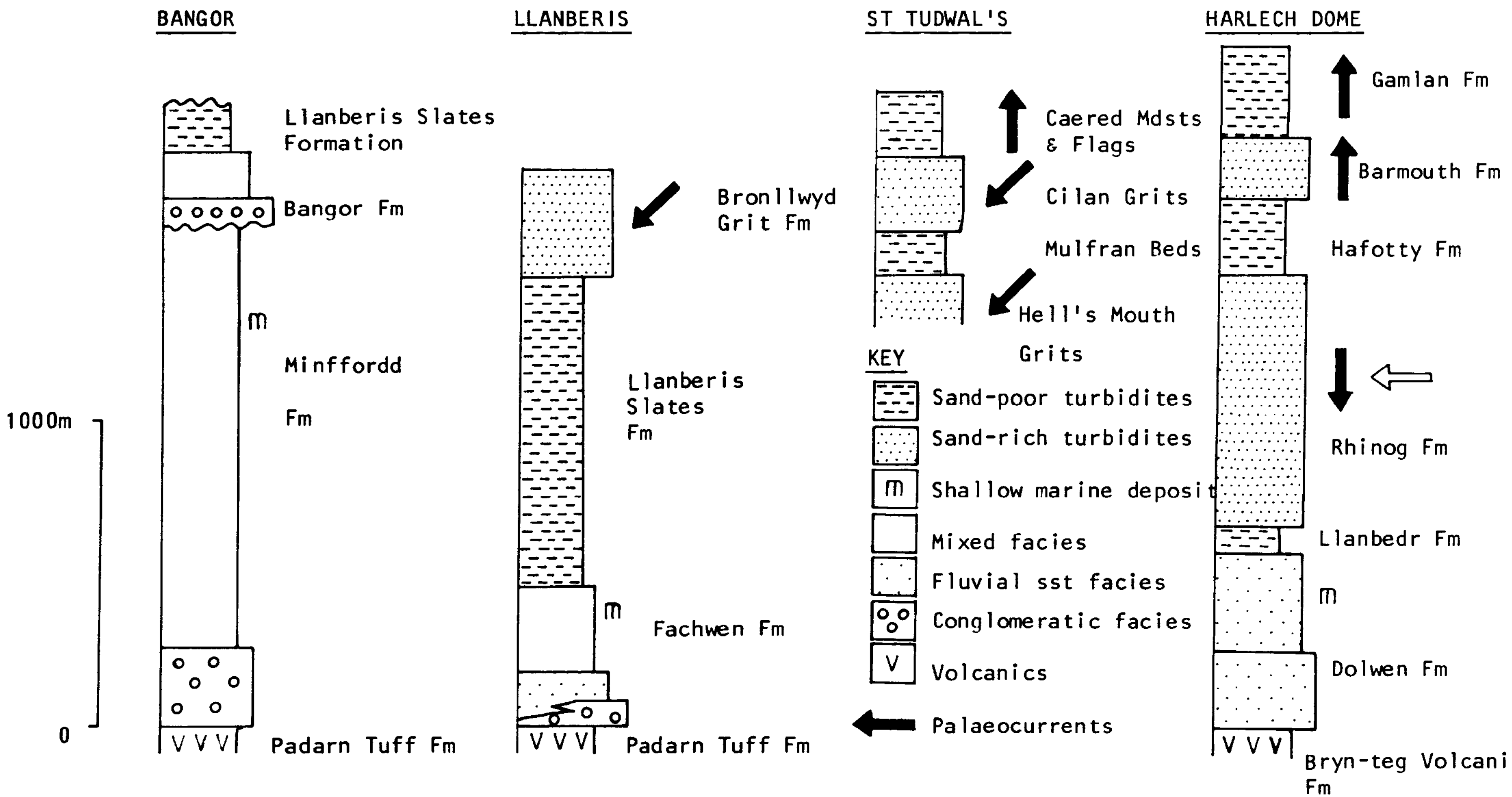


FIG 8.3 Summary diagram of facies, palaeocurrents and successions from the Lower and Middle Cambrian of North Wales.



REFERENCES

REFERENCES.

AALTO, K.R. 1976. Sedimentology of a melange: Franciscan of Trinidad, California. *Journal of Sedimentary Petrology*, 46, 913-29.

ALLEN, J.R.L. 1965. A review of the origin and characteristics of recent alluvial sediments. *Sedimentology*, 5, 89-191.

ALLEN, J.R.L. 1970. The sequence of sedimentary structures in turbidites with special reference to dunes. *Scottish Journal of Geology*, 6, 146-161.

ALLEN, J.R.L. 1971. Mixing at Turbidity Current heads and its geological implications. *Journal of Sedimentary Petrology*, 41, 97-113.

ALLEN, J.R.L. 1974. Reaction, Relaxation and Lag in natural sedimentary systems: general principles, examples and lessons. *Earth Science Reviews*, 10, 263-342.

ALLEN, J.R.L. 1977. The possible mechanics of convolute lamination in graded sand beds. *Journal of the Geological Society, London*, 134, 19-31.

ALLEN, J.R.L. 1982. *Sedimentary Structures: Their Character and Physical Basis.* Volumes 1 and 2, Elsevier, Amsterdam.

ALLEN, J.R.L. & FRIEND, P.F. 1976. Relaxation time of dunes in decelerating aqueous flows. *Journal of the Geological Society, London*, 132, 17-26.

ALLEN, P.M., COOPER, D.C., FUGE, R. & REA, W.J. 1976. Geochemistry and relationships to mineralisation of some igneous rocks from the Harlech Dome, North Wales. *Transactions of the Institution of Mining and Metallurgy*, Section B, 85, 100-8.

ALLEN, P.M. & EASTERBROOK, G.D. 1978. Mineralised breccia pipe and other intrusion breccias in the Harlech Dome, North Wales. *Transactions of the Institution of Mining and Metallurgy*, Section B, 87, 157-61.

ALLEN, P.M. & JACKSON, A.A. 1978. Bryn-teg borehole, North Wales. *Bulletin of the Geological Survey of Great Britain*, 61.

ALLEN, P.M. & JACKSON, A.A. 1985. Geology of the country around

Harlech. *Memoir of the British Geological Survey*, Sheet 135 with part of 149.

ALLEN, P.M., JACKSON, A.A. & RUSHTON, A.W.A. 1981. The Stratigraphy of the Mawddach Group in the Cambrian succession of North Wales. *Proceedings of the Yorkshire Geological Society*, 43, 295-329.

ANDERTON, R. 1980. Did Iapetus start to open during the Cambrian? *Nature*, 286, 706-8.

ANDERTON, R., BRIDGES, P.H., LEEDER, M.R & SELLWOOD, B.W. 1979. *A Dynamic Stratigraphy of the British Isles.* George Allen & Unwin, London.

ANDREW, A.R. 1910. Geology of the Dolgelley gold belt, North Wales. *Geological Magazine*, 7, 159-71, 201-11, & 261-71.

BAILEY, E.B. 1930. New light on sedimentation and tectonics. *Geological Magazine*, 67, 77-92.

BAKER, J.W. 1973. A marginal late Proterozoic ocean basin in the Welsh region. *Geological Magazine*, 110, 447-55.

BARBER, A.J. & MAX, M.D. 1979. A new look at the Mona Complex, (Anglesey, North Wales). *Journal of the Geological Society, London*, 136, 407-32.

BARBER, A.J., MAX, M.D. & BRUCK, P.M. 1981. Geologists' Association-Irish Geological Association; Field meeting in Anglesey and southeastern Ireland, 4-11 June 1977. *Proceedings of the Geologists' Association, London*, 92, 269-91.

BASSETT, D.A. 1963. The Welsh Palaeozoic geosyncline: a review of recent work on stratigraphy and sedimentation. *In*: STEWART, F.H. & JOHNSON, M.R.W. (eds.) *The British Caledonides*. Oliver and Boyd, Edinburgh, 35-69.

BASSETT, D.A. & WALTON, E.K. 1960. The Hell's Mouth Grits: Cambrian greywackes in St. Tudwal's peninsula, North Wales. *Quarterly Journal of the Geological Society, London*, 116, 85-110.

BASSETT, M.G., OWENS, R.M. & RUSHTON, A.W.A. 1976. Lower Cambrian fossils from the Hell's Mouth Grits, St. Tudwal's, North Wales. *Journal of the Geological Society, London*, 132, 623-44.

BATES, D.E.B. 1975. Slaty cleavage associated with sandstone dykes in the Harlech Dome, North Wales. *Geological Journal*, 10, 167-75.

BECKINSALE, R.D., EVANS, J.A., THORPE, R.S., GIBBONS, W. & HARMON, R.S. 1984. Rb-Sr whole-rock isochron ages, del ^{18}O values and geochemical data for the Sarn Igneous Complex and the Parwyd gneisses of the Mona Complex of Llyn, N. Wales. *Journal of the Geological Society, London*, 141, 701-9.

BECKINSALE, R.D. & THORPE, R.S. 1979. Rubidium-strontium whole-rock isochron evidence for the age of metamorphism and magmatism in the Mona Complex of Anglesey. *Journal of the Geological Society, London*, 136, 433-39.

BENNETT, M.A. 1987. Genesis and diagenesis of the Cambrian manganese deposits, Harlech, North Wales. *Geological Journal*, 22, 7-18.

BEVINS, R.E. & ROWBOTHAM, G. 1983. Low grade metamorphism within the Welsh sector of the paratectonic Caledonides. *Geological Journal*, 18, 141-67.

BINSTOCK, J.L.H. 1977. *Petrology and sedimentation of the Cambrian manganese-rich sediments of the Harlech Dome, North Wales*. Ph.D thesis, Harvard University.

BLAKE, J.F. 1888. On the Cambrian and associated rocks in NW. Caernarvonshire. *Quarterly Journal of the Geological Society, London*, 44, 271-90.

BLAKE, J.F. 1892. On the rocks mapped as Cambrian in Caernarvonshire. *Quarterly Journal of the Geological Society, London*, 48, 243-62.

BLAKE, J.F. 1893. On the felsites and conglomerates between Bethesda and Llanllyfni, north Wales. *Quarterly Journal of the Geological Society, London*, 49, 441-65.

BLAKE, J.F. 1898. A revindication of the Llanberis unconformity. *Geological Magazine*, 5, 169-78, 214-26.

BONNEY, T.G. 1879. On the quartz-felsite and associated rocks at the base of the Cambrian series in north-western Caernarvonshire. *Quarterly Journal of the Geological Society, London*, 35, 309-20.

BONNEY, T.G. & RAISON, C.A. 1894. On the relations of some of the older fragmental rocks in NW. Caernarvonshire. *Quarterly Journal of the Geological Society, London*, 50, 578-602.

BOSE, P.K. 1983. A reappraisal of the conditions of deposition of the Maentwrog beds (Upper Cambrian) of Porth Ceiriad, north Wales. *Geological Magazine*, 120, 73-80.

BOUMA, A.H. 1962. *Sedimentology of some flysch deposits*. Elsevier, Amsterdam.

BOUMA, A.H. 1972. Fossil contourites in lower Niesenflysch, Switzerland. *Journal of Sedimentary Petrology*, 42, 917-21.

BOUMA, A.H. & TREADWELL, T.K. 1975. Deep-sea Dune-like features. *Marine Geology*, 19, M53-59.

BRASIER, M.D. 1980. The lower Cambrian transgression and glauconite-phosphate facies in western Europe. *Journal of the Geological Society, London*, 137, 695-703.

BRASIER, M.D. 1982. Sea level changes, facies changes and the late Precambrian-early Cambrian evolutionary explosion. *Precambrian Research*, 17, 105-23.

BRASIER, M.D. & HEWITT, R.A. 1979. Environmental setting of fossiliferous rocks from the uppermost Proterozoic-lower Cambrian of central England. *Palaeogeography, Palaeoclimatology & Palaeoecology*, 27, 35-57.

BRASIER, M.D., HEWITT, R.A. & BRASIER, C.J. 1978. On the late Precambrian-early Cambrian Hartshill Formation of Warwickshire. *Geological Magazine*, 115, 21-36.

BRENCHLEY, P.J. 1969. Origin of matrix in Ordovician greywackes, Berwyn hills, north Wales. *Journal of Sedimentary Petrology*, 39, 1297-1301.

BRENCHLEY, P.J. 1985. Storm influenced sandstone beds. *Modern Geology*, 9, 369-96.

BRENCHLEY, P.J., HARPER, J.C & SKEVINGTON, D. 1967. Lower Ordovician shelly and graptolitic faunas from south-eastern Ireland. *Proceedings of the Royal Irish Academy*, 65, Section B, 385-90.

BREWER, J.A., MATTHEWS, D.H., WARNER, M.C., HALL, J., SMYTHE, D.K. & WHITTINGTON, R.J. 1983. BIRPS deep seismic reflection studies of the British Caledonides. *Nature*, 305, 206-10.

BRUCK, P.M., POTTER, T.L. & DOWNIE, C. 1974. The Lower Palaeozoic stratigraphy of the northern part of the Leinster Massif. *Proceedings of the Royal Irish Academy*, 74, Section B, 75-84.

BUCHBINDER, L.G., MAGARITZ, M. & GOLDBERG, M. 1984. Stable isotope study of karstic-related dolomitisation: Jurassic rocks from the Coastal Plain, Israel. *Journal of Sedimentary Petrology*, 54, 236-56.

BUSBY-SPERA, C. 1985. A sand-rich submarine fan in the Lower Mesozoic Mineral King Caldera Complex, Sierra Nevada, California. *Journal of Sedimentary Petrology*, 55, 376-91.

CALVERT, S.E. & PRICE, N.B. 1972. Diffusion and reaction profiles of dissolved manganese in the pore waters of marine sediments. *Earth & Planetary Science Letters*, 16, 245-49.

CAS, R. 1979. Mass-flow arenites from a Palaeozoic interarc basin, New South Wales, Australia: mode and environment of emplacement. *Journal of Sedimentary Petrology*, 49, 29-44.

CATTERMOLE, P. & JONES, A. 1970. The geology of the area around Mynydd Mawr, Nantlle, Caernarvonshire. *Geological Journal*, 7, 111-28.

CATTERMOLE, P. & ROMANO, M. 1981. *Lleyn Peninsula.* Geologists' Association, London, Guide No. 39.

CHIPPING, D.H. 1972. Sedimentary structure and environment of some thick sandstone beds of turbidite type. *Journal of Sedimentary Petrology*, 42, 587-95.

COBBOLD, E.S. 1927. The stratigraphy and geological structure of the Cambrian area of Comley (Shropshire). *Quarterly Journal of the Geological Society, London*, 83, 551-73.

COCKS, L.R.M. & FORTEY, R.A. 1982. Faunal evidence for oceanic separation in the Palaeozoic of Britain. *Journal of the Geological Society, London*, 139, 465-78.

COGNE, J. & LEFORT, L.P. 1985. The Ligerian Orogeny: a proto-Variscan event, related to the Siluro-Devonian evolution of the Tethys I Ocean, *In*: GEE, D.G. & STURT, B.A. (eds.) *The Caledonide Orogen: Scandinavia and related areas.* John Wiley & sons, Chichester. Volume 2, 1185–93.

COGNE, J. & WRIGHT, A.E. 1980. L'Orogene cadomien; vers un essai d'interpretation paleogeodynamique unitaire des phenomenes orogeniques fini-precambriens d'Europe moyenne et occidentale, et leur signification a l'origine de la croite et du mobilisme varisque puis alpin. *In*: COGNE, J. *et al.* (eds.) *Geologie de l'Europe du Precambrien aux bassins sedimentaires post-hercyniens.* Fr. Bur. Rech. Geol. Minieres. Mem. 108, 29–55.

COLEMAN, J.M. 1969. Brahmaputra River: channel processes and sedimentation. *Sedimentary Geology*, 3, 129–239.

COLEMAN, M., FLEET, A. & DONSON, P. 1982. Preliminary studies of manganese-rich carbonate nodules from Leg 68, Site 503, Eastern Equatorial Pacific. *Initial Reports of the Deep Sea Drilling Project*, 68, 481–89.

COPE, J.C.W. & GIBBONS, W. 1987. New evidence for the relative age of the Ercall Granophyre and its bearing on the Precambrian-Cambrian boundary in southern Britain. *Geological Journal*, 22, 53–60.

COWIE, J.W. 1971. Lower Cambrian faunal provinces. *In*: MIDDLEMISS, F.A., RAWSON, P.F. & NEWALL, G. (eds.), *Faunal Provinces in Time and Space*, Geological Journal Special Issue, No. 4, 31–46. Seel House Press, Liverpool.

COWIE, J.W., RUSHTON, A.W.A. & STUBBLEFIELD, C.J. 1972. *Correlation of the Cambrian Rocks in the British Isles.* Geological Society, London, Special Report, No. 2.

COX, A.H., GREEN, J.F.N., JONES, O.T. & PRINGLE, J. 1930. Geology of the St. David's district, Pembrokeshire. *Proceedings of the Geologists' Association, London*, 41, 241–73.

CRAIG, G.Y. & WALTON, E.K. 1962. Sedimentary structures and palaeocurrent directions from the Silurian rocks of Kirkcudbrightshire. *Transactions of the Geological Society of Edinburgh*, 19, 100–19.

CRAIG, H. 1957. Isotopic standards for carbon amd oxygen correction factors for mass spectrometric analysis of carbon dioxide. *Geochimica et Cosmochimica Acta*, 12, 133–39.

CRIMES, T.P. 1966. The relative age of some concretions in Cambrian sediments of St Tudwal's peninsula, north Wales. *Geological Journal*, 5, 33-42.

CRIMES, T.P. 1969. *The stratigraphy, structure and sedimentology of some Precambrian and Cambro-Ordovician rocks bordering the southern Irish Sea.* Ph.D thesis, University of Liverpool.

CRIMES, T.P. 1970a. A facies analysis of the Cambrian of Wales. *Palaeogeography, Palaeoclimatology, Palaeoecolology*, 7, 113-70.

CRIMES, T.P. 1970b. The significance of trace fossils in sedimentology, stratigraphy and palaeoecology with examples from Lower Palaeozoic strata. *In:* CRIMES, T.P. & HARPER, J.C. (eds.) *Trace Fossils.* Geological Journal Special Publication No 3, 101-26. Seel House Press, Liverpool.

CRIMES, T.P. 1974. Colonisation of the early ocean floor. *Nature*, 248, 328-30.

CRIMES, T.P. 1975. The stratigraphical significance of trace fossils. *In:* FREY, R.W. (ed.), *The Study of Trace Fossils.* Springer, New York, 109-30.

CRIMES, T.P. 1976. Trace fossils from the Bray Group (Cambrian) of Howth, County Dublin. *Bulletin of the Geological Survey of Ireland*, 2, 53-67.

CRIMES, T.P. 1977. Trace fossils of an Eocene deep-sea sand fan, N. Spain. *In:* CRIMES, T.P. & HARPER, J.C. (eds.) *Trace Fossils 2.* Geological Journal Special Publication No 9, 71-90. Seel House Press, Liverpool.

CRIMES, T.P. & CROSSLEY, J.D. 1968. The stratigraphy, sedimentology, ichnology and structure of the Lower Palaeozoic rocks of part of northeastern County Wexford. *Proceedings of the Royal Irish Academy*, Section B, 67, 185-215.

CRIMES, T.P., GOLDRING, R., HOMEWOOD, P., VAN STUIJVENBERG, J. & WINKLER, W. 1981. Trace fossil assemblages of deep-sea fan deposits, Gurnigel and Schlieren flysch (Cretaceous-Eocene), Switzerland. *Eclogae Geologicae Helvetiae*, 74, 953-95.

CRIMES, T.P. & OLDERSHAW, M.A. 1967. Palaeocurrent determinations by magnetic fabric measurements on the Cambrian rocks of St. Tudwal's peninsula, North Wales. *Geological Journal*, 5, 217-32.

CRIMES, T.P. & SLY, P.G. 1964. Implications of certain sedimentary structures within the Cambrian succession of St Tudwal's peninsula, SW Caernarvonshire. *Nature*, 204, 174-75.

CUMMINS, W.A. 1962. The greywacke problem. *Liverpool & Manchester Geological Journal*, 3, 51-72.

DEARNLEY, R. 1967. Ignimbrites from the Uriconian and Arvonian. *Bulletin of the Geological Survey of Great Britain*, 24, 1-6.

DE BETHUNE, S. 1972. Etude a la microsonde electronique des mineraux metamorphiques du Cambrien de Harlech. (Pays de Galles). *Bulletin de la Societe Belge de Geologie, Paleontologie et d'Hydrologie*, 81, 255-77.

DEINES, P. 1970. Mass spectrometer correction factors for the determination of small isotopic variations of carbon and oxygen. *International Journal of Mass Spectrometry and Ion Physics*, 4, 283-95.

DEMAISON, G.J. & MOORE, G.T. 1980. Anoxic environments and oil source bed genesis. *Bulletin of the American Association of Petroleum Geologists*, 64, 1179-1209.

DEWEY, J.F. 1969. Evolution of the Appalachian-Caledonian orogen. *Nature*, 222, 124-9.

DICKINSON, W.R. & SUCZEK, C.A. 1979. Plate tectonics and sandstone compositions. *Bulletin of the American Association of Petroleum Geologists*, 63, 2164-2182.

DOTT, R.H. & BOURGEOIS, J. 1982. Hummocky stratification: significance of its variable bedding sequences. *Bulletin of the Geological Society of America*, 93, 663-80.

DRUITT, T.H. & SPARKS, R.S.J. 1982. A proximal ignimbrite breccia facies on Santorini, Greece. *Journal of Volcanology and Geothermal Research*, 13, 147-71.

DZULYNSKI, S., KSIAZKIEWICZ, M. & KUENEN, Ph.H. 1959. Turbidites in flysch of the Polish Carpathian mountains. *Bulletin of the Geological Society of America*, 70, 1089-118.

DZULYNSKI, S. & WALTON, E.K. 1965. *Sedimentary features of* *flysch and greywackes*. Elsevier, Amsterdam.

EASTWOOD, T., WHITEHEAD, T.H. & ROBERTSON, T. 1925. The geology of the country around Birmingham. *Memoir of the Geological Survey of Great Britain*, Sheet 168.

EVANS, C.D.R. 1968. *Geological succession and structure of the area east of Bethesda*. Ph.D thesis, University of Aberystwyth.

FAUGERES, J.C., GONTHIER, E. & STOW, D.A.V. 1984. Contourite drift moulded by deep Mediterranean outflow. *Geology*, 12, 296-300.

FEARNSIDES, W.G. 1910a. North and Central Wales. *Geology in the Field*, 2, 786-825.

FEARNSIDES, W.G. 1910b. The Tremadoc slates and associated rocks of south-east Carnarvonshire. *Quarterly Journal of the Geological Society, London*, 66, 142-88.

FEARNSIDES, W.G. & DAVIES, W. 1944. The geology of Deudraeth, the country between Traeth Mawr and Traeth Bach, Merioneth. *Quarterly Journal of the Geological Society, London*, 99, 247-276.

FISHER, R.V. 1971. Features of coarse grained, high concentration fluids and their deposits. *Journal of Sedimentary Petrology*, 41, 916-27.

FISHER, R.V. 1983. Flow transformations in sediment gravity flows. *Geology*, 11, 273-74.

FOLK, R.L. 1974. *Petrology of sedimentary rocks*. Hemphills Publishing Company, Austin.

FOSTER, H. 1968. *The glaciation of the Harlech Dome*. Ph.D. thesis, University of London.

FRIEDMAN, I. & O'NEILL, J.R. 1977. Compilation of stable isotope fractionation factors of geochemical interest. *United States Geological Survey Professional Papers*, 440-KK.

GARDINER, P.R.R. & VANGUESTAINE, M. 1971. Cambrian and Ordovician microfossils fron south-east Ireland and their implications. *Bulletin of the Geological Survey of Ireland*, 1, 163-210.

GEIKE, A. 1891. Anniversary address of the president. *Proceedings of the Geological Society, London*, 47, 48-162.

GIBBONS, W. 1983a. Stratigraphy, subduction and strike-slip faulting in the Mona Complex of North Wales- a review. *Proceedings of the Geologists' Association, London,* 94, 147-63.

GIBBONS, W. 1983b. The Monian "Penmynydd zone of metamorphism" in Llyn, North Wales. *Geological Journal,* 18, 21-41.

GIBBONS, W. 1984. The Precambrian basement of England and Wales. *Proceedings of the Geologists' Association, London.* 95, 387-89.

GLASBY, G.P. 1974. A geochemical study of the manganese ore deposits of the Harlech Dome, North Wales. *Journal of Earth Science, Leeds Geological Association,* 8, 445-50.

GOKCEN, S.L. & KELLING, G. 1983. The Palaeogene Yamak sand-rich submarine fan Complex, Haymana Basin, Turkey. *Sedimentary Geology,* 34, 219-243.

GOLD, P.B. 1987. Textures and geochemistry of authigenic albite from Miocene sandstones, Louisiana Gulf Coast. *Journal of Sedimentary Petrology,* 57, 353-62.

GREEN, A.H. 1885. Note on a section near Llanberis. *Quarterly Journal of the Geological Society, London,* 41, 74-9.

GREENLY, E. 1897. Incipient metamorphism in the Harlech grits. *Transactions of the Geological Society of Edinburgh* 7, 254-8.

GREENLY, E. 1919. Geology of Anglesey. *Memoir of the Geological Survey of Great Britain,* volumes 1 and 2.

GREENLY, E. 1923. Further researches on the succession and metamorphism in the Mona Complex of Anglesey. *Quarterly Journal of the Geological Society, London,* 79, 334-51.

GREENLY, E. 1944. The Cambrian rocks of Arvon. *Geological Magazine,* 81, 170-5.

GREENLY, E. 1945. The Arvonian rocks of Arvon. *Quarterly Journal of the Geological Society, London,* 100, 269-8.

GROOM, T.T. 1902. The sequence of the Cambrian and associated beds of the Malvern Hills. *Quarterly Journal of the Geological Society, London,* 58, 89-149.

HAGEMANN, F. 1957. On the petrography of the Silurian slates from Hadeland, Norway. *Norsk Geologisk Tidsskrift*, 37, 229-246.

HAMAR, G. 1967. *Platysolenites antiquissimus* Eichw. (Vermes) from the Lower Cambrian of northern Norway. *Norges Geologiske Undersokelse*, 249, 89-95.

HAWKINS, T.R.W. 1985. Influence of geological structure on slope stability in the Maentwrog Formation, Harlech Dome, North Wales. *Proceedings of the Geologists' Association, London*, 96, 289-304.

HAWKINS, T.R.W & JONES, F.G. 1981. Fold and cleavage development within Cambrian metasediments of the Vale of Ffestiniog, North Wales. *Geological Journal*, 16, 65-84.

HEIN, F.J. 1982. Depositional mechanisms of deep-sea coarse clastic sediments, Cap Enrage Formation, Quebec. *Canadian Journal of Earth Science*, 19, 267-287.

HEIN, F.J. & WALKER, R.G. 1982. The Cambro-Ordovician Cap Enrage Formation, Quebec, Canada: conglomeratic deposits of a braided submarine channel with terraces. *Sedimentology*, 29, 309-29.

HELM, D.G. & PICKERING, K.T. 1985. The Jersey Shale Formation- a late Precambrian deep-water siliciclastic system, Jersey, Channel Islands. *Sedimentary Geology*, 43, 43-66.

HENDRY, H.E. 1972. Breccias deposited by mass flow in the Breccia Nappe of the French pre-Alps. *Sedimentology*, 18, 277-92.

HENDRY, H.E. 1978. Cap des Rosiers Formation at Grosses Roches, Quebec- deposits of the mid-fan region on an Ordovician submarine fan. *Canadian Journal of Earth Science*, 15, 1472-88.

HICKS, H. 1878. On some Pre-Cambrian (Dimetian and Pebidian) rocks in Caernarvonshire. *Quarterly Journal of the Geological Society, London*, 34, 147-52.

HISCOTT, R.N. & MIDDLETON, G.V. 1979. Depositional mechanics of thick-bedded sandstones at the base of a sandstone slope, Tourelle Formation (Lower Ordovician), Quebec, Canada. *In*: DOYLE, L.J. & PILKEY, O.H. (eds.) *Geology of Continental Slopes*. Society of Economic Paleontologists &

Mineralogists Special Publication, 27, 307-26.

HOWELL, B.F. & STUBBLEFIELD, C.J. 1950. A revision of the fauna of the North Welsh "Conocoryphe viola" beds implying Lower Cambrian age. *Geological Magazine*, 87, 1-16.

HOWELLS, M.F., REEDMAN, A.J. & LEVERIDGE, B.E. 1985. *Geology of the country around Bangor, explanation of the 1:50,000 sheet*. British Geological Survey Sheet 106.

HUBERT, C., LAJOIE, J. & LEONARD, M.A. 1970. Deep sea sediments in the lower Palaeozoic Quebec Supergroup. *In*: LAJOIE, J. (ed.) *Flysch Sedimentology in North America*. Geological Association of Canada, Special Paper, 7, 103-125.

HUGHES, E.W. 1917. On the geology of the district from Cil-y-Coed to the St. Anne's-Llanllyfni ridge (Caernarvonshire). *Geological Magazine*, 54, 12-25.

IRWIN, H., CURTIS, C. & COLEMAN, M. 1977. Isotopic evidence for source of diagenetic carbonates formed during burial of organic-rich sediments. *Nature*, 269, 209-213.

JAMES, N.P. & CHOQUETTE, P.W. 1983. Diagenesis 6. Limestones, the seafloor diagenetic environment. *Geoscience Canada*, 10, 162-179.

JOHNSON, B.A. & WALKER, R.G. 1979. Paleocurrents and depositional environments of deep water conglomerates in the Cambro-Ordovician Cap Enrage Formation, Quebec Appalachians. *Canadian Journal of Earth Science*, 16, 1375-87.

JOHNSON, T.C., CARLSON, T.W. & EVANS, J.E. 1980. Contourites in Lake Superior. *Geology*, 8, 437-41.

JONES, B. 1933. Geology of the Fairbourne-Llwyngwril district, Merioneth. *Quarterly Journal of the Geological Society, London*, 89, 145-171.

JONES, O.T. 1938. On the evolution of a geosyncline. *Quarterly Journal of the Geological Society, London*, 94, lx-cx.

JONES, O.T. 1955. The geological evolution of Wales and the adjacent regions. *Quarterly Journal of the Geological Society, London*, 111, 323-51.

KASTNER, M. & SIEVER, R. 1979. Low temperature feldspars in sedimentary rocks. *American Journal of Science*, 279, 435-79.

KELLING, G. 1964. The turbidite concept in Britain. *In*: BOUMA, A.H. & BROUWER, A. (eds.) *Turbidites*. Elsevier, Amsterdam, 75-92.

KNILL, J.L. 1959. Axial and marginal sedimentation in geosynclinal basins. *Journal of Sedimentary Petrology*, 29, 317-25.

KOKELAAR, B.P. 1977. *The Igneous history of the Rhobell Fawr Area, Merioneth, North Wales.* Ph.D thesis, University of Aberystwyth.

KOKELAAR, B.P. 1979. Tremadoc to Llanvirn volcanism on the SE. side of the Harlech Dome (Rhobell Fawr), North Wales. *In*: HARRIS, A.L., HOLLAND, C.H. & LEAKE, B.E. (eds.) *The Caledonides of the British Isles- reviewed*. Special Publication of the Geological Society, London, No. 8, 591-6.

KOKELAAR, B.P., HOWELLS, M.F., BEVINS, R.E., ROACH, R.A. & DUNKLEY, P.N. 1984. The Ordovician marginal basin of Wales. *In*: KOKELAAR, B.P. & HOWELLS, M.F. (eds.) *Marginal Basin Geology*. Special Publication of the Geological Society, London, No. 16, 245-69.

KOPSTEIN, F.W. 1954. *Graded Bedding of the Harlech Dome*. Ph.D thesis, University of Groningen.

KUENEN, Ph.H. 1953. Graded bedding with observations on Lower Palaeozoic rocks of Britain. *Verhandelingen der Koninklijke Nederlandse Akademie van Wetenschappen afd. Natuurkunde. Erste Reeks Deel XX*, 3, North Holland Publishing Company.

KUENEN, Ph.H. & MIGLIORINI, C.I. 1950. Turbidity currents as a cause of graded bedding. *Journal of Geology*, 58, 91-127.

LAJOIE, J. 1979. Origin of megarhythms in flysch sequences of the Quebec Appalachians. *Canadian Journal of Earth Science*, 16, 1518-23.

LAND, L.S. & MILLIKEN, K.L. 1981. Feldspar diagenesis in the Frio Formation, Brazoria County, Texas Gulf Coast. *Geology*, 9, 314-18.

LEEDER, M.R. 1982. Upper Palaeozoic basins of the British Isles- Caledonide inheritance versus Hercynian plate margin processes. *Journal of the Geological Society, London*, 137, 139-56.

LEGGETT, J.K. 1980. British Lower Palaeozoic black shales and their palaeo-oceanographic significance. *Journal of the Geological Society, London*, 137, 139-56.

LEGGETT, J.K., MCKERROW, W.S. & SOPER, N.J. 1983. A model for the crustal evolution of southern Scotland. *Tectonics*, 2, 187-210.

LINK, M.H. & NILSEN, T.H. 1980. The Rocks Sandstone, an Eocene sand-rich deep-sea fan deposit, northern Santa Lucia Range, California. *Journal of Sedimentary Petrology*, 50, 583-602.

LINK, M.H. & WELTON, J.E. Sedimentology and reservoir potential of Matilija Sandstone: an Eocene sand-rich deep-sea fan and shallow marine complex, California. *American Association of Petroleum Geologists, Bulletin*, 66, 1514-34.

LIPMAN, P.W. 1975. Evolution of the Platero Caldera complex and related volcanic rocks of south-eastern San Juan Mountains, south-western Colorado. *United States Geological Survey Professional Papers*, 852.

LOCKLEY, M.G. & WILCOX, C.J. 1979. A Lower Cambrian brachiopod from the Harlech Dome. *Geological Magazine*, 116, 63-4.

LOWE, D.R. 1976. Grain flow and grain flow deposits. *Journal of Sedimentary Petrology*, 46, 188-99.

LOWE, D.R. 1982. Sediment gravity flows II: depositional models with special reference to the deposits of high density turbidity currents. *Journal of Sedimentary Petrology*, 52, 279-97.

LOWE, D.R. & LOPICCOLO, R.D. 1974. The characteristics and origins of dish and pillar structures. *Journal of Sedimentary Petrology*, 44, 484-501.

LYNN, D.C. & BONATTI, E. 1965. Mobility of manganese in diagenesis of deep sea sediments. *Marine Geology*, 3, 457-74.

MCCREA, J.M. 1950. On the isotopic chemistry of carbonates and a palaeotemperature scale. *Journal of Chemistry & Physics*, 18, 849-57.

MCKENZIE, D. 1978. Some remarks on the development of sedimentary basins. *Earth & Planetary Science Letters*, 40, 25-32.

MCKERROW, W.S. LEGGETT, J.K. & EALES, M.H. 1977. Imbricate thrust model of the Southern Uplands of Scotland. *Nature*, 267, 237-39.

MALTMAN, A.J. 1975. Ultramafic rocks in Anglesey: their non-tectonic emplacement. *Journal of the Geological Society, London*, 131, 593-605.

MATLEY, C.A. 1928. The Pre-Cambrian complex and associated rocks of SW. Lleyn (Caernarvonshire). *Quarterly Journal of the Geological Society, London*, 84, 440-504.

MATLEY, C.A., NICHOLAS, T.C. & HEARD, A. 1939. Summer field meeting to the western part of the Lleyn Peninsula. *Proceedings of the Geologists' Association, London*, 50, 83-100.

MATLEY, C.A. & WILSON, T.S. 1946. The Harlech Dome, north of the Barmouth estuary. *Quarterly Journal of the Geological Society, London*, 102, 1-40.

MATTHEWS, S.C. & COWIE, J.W. 1979. Early Cambrian transgression. *Journal of the Geological Society, London*, 136, 133-35.

MAX, M.D. 1975. Precambrian rocks of southeast Ireland. *In*: HARRIS, A.L., SHACKLETON, R.M., WATSON, J., DOWNIE, C., HARLAND, W.B. & MOORBATH, S. (eds.) *A correlation of the Precambrian rocks in the British Isles*. Geological Society, London, Special Report 6, 97-101.

MIDDLETON, G.V. & HAMPTON, M.A. 1976. Subaqueous sediment transport and deposition by sediment gravity flows. *In*: STANLEY, D.J. & SWIFT, D.J.P. (eds.) *Marine sediment transport and environmental management*. John Wiley, New York, 197-218.

MIDDLETON, G.V. & SOUTHARD, J.V. 1977. *Mechanics of sediment movement*. Society of Economic Paleontologists & Mineralogists, short course No. 3.

MOHR, P.A. 1955. A suspected spilitic rock from the Lower Cambrian Manganese Shale Group of the Harlech Dome, North Wales. *Geological Magazine*, 99, 265-72.

MOHR, P.A. 1956a. Trace element distribution in a garnet-chlorite rock from Foel Ddu, near Harlech, Merionethshire. *Mineralogical Magazine*, 31, 319-327.

MOHR, P.A. 1956b. A geochemical study of the Lower Cambrian Manganese ore of the Harlech Dome, North Wales. *In: Symposium del Manganeso XX Congreso Geologico Internacional*, 273-89.

MOHR, P.A. 1959. A geochemical study of the shales of the Lower Manganese Shale Group of the Harlech Dome, North Wales. *Geochimica et Cosmochimica Acta*, 17, 186-200.

MOHR, P.A. 1964. Genesis of the Cambrian manganese carbonate rocks of North Wales. *Journal of Sedimentary Petrology*, 34, 819-29.

MOHR, P.A. & ALLEN, R. 1965. Further considerations on the deposition of Middle Cambrian Manganese carbonate beds of Wales and Newfoundland. *Geological Magazine*, 102, 328-37.

MORRIS, T.O. & FEARNSIDES, W.G. 1926. The stratigraphy and structure of the Cambrian Slate Belt of Nantlle (Caernarvonshire). *Quarterly Journal of the Geological Society, London*, 82, 250-303.

MUIR, M.D., BLISS, G.M., GRANT, P.R. & FISHER, M.J. 1979. Palaeontological evidence for the age of some supposedly Precambrian rocks in Anglesey, North Wales. *Journal of the Geological Society, London*, 136, 61-64.

MUTTI, E. 1979. Turbidites et cones sous-marins profonds. *In:* HOMEWOOD, P. (ed.) *Sedimentation detritique.* 3eme cycle Romand Sciences de la terre, Fribourg, 1, 353-419.

MUTTI, E., FONNESU, F., RAMPONE, G. & SONNINO, M. 1981. Channel-fill and associated overbank deposits in the Eocene Hecho Group, Ainsa-Boltana region (South-central Pyrenees). *International Association of Sedimentologists, 2nd. European Meeting Bologna. 1981. Abstracts.* 113-6.

MUTTI, E. & NILSEN, T.H. 1981. Significance of intraformational rip-up clasts in deep-sea fan deposits. *International Association of Sedimentologists 2nd. European Meeting Bologna. 1981. Abstracts.* 117-9.

MUTTI, E. & RICCI LUCCHI, F. 1972. Le torbiti dell'Appen. settentionale intro. all'analisi di facies. *Memorie della*

Societa Geologica Italiana, 161-199. English translation in *International Geological Review*, for 1978, 20, 125-66.

MUTTI, E., RICCI LUCCHI, F., SEGURET, M. & ZANZUCCHI, G. 1984. Seismoturbidites: a new group of resedimented deposits. *Marine Geology*, 55, 103-116.

MUTTI, E. & SONNINO, M. 1981. Compensation cycles: a diagnostic feature of turbidite sandstone lobes. *International Association of Sedimentologist 2nd. European Meeting Bologna. 1981. Abstracts.* 120-3.

NICHOLAS, T.C. 1915. Geology of the St. Tudwal's Peninsula (Caernarvonshire). *Quarterly Journal of the Geological Society, London*, 71, 83-143.

NICHOLAS, T.C. 1916. Notes on the trilobite fauna of the Middle Cambrian of St. Tudwal's. *Quarterly Journal of the Geological Society, London*, 71, 83-141.

NILSEN, T.H. & CLARKE, S.H. 1975. Sedimentation and Tectonics in the Early Tertiary Continental Borderland of Central California. *United States Geological Survey Professional Papers*, 925.

NORMARK, W.R., HESS, G.R., STOW, D.A.V. & BOWEN, A.J. 1980. Sediment waves on the Monterey Fan levee: a preliminary physical interpretation. *Marine Geology*, 37, 1-18.

NUTT, M.J.C. & SMITH, E.G. 1981. Transcurrent faulting and the anomalous position of pre-Carboniferous Anglesey. *Nature*, 290, 492-5.

OKADA, H. 1966. Non-greywacke "turbidite" sandstones in the Welsh geosyncline. *Sedimentology*, 7, 211-32.

OKADA, H. 1967. Composition and cementation of some Lower Palaeozoic grits in Wales. *Memoir of the Faculty of Science, Kyushu University (Series D) Geology*, 18, 261-76.

O'NIONS, R.K., OXBURGH, E.R., HAWKESWORTH, C.J. & MACINTYRE, R.M. 1973. New isotopic and stratigraphic evidence on the age of the Ingletonian: probable Cambrian of northern England. *Journal of the Geological Society, London*, 129, 445-52.

OPIK, A.A. 1956. Cambrian (Lower Cambrian) of Estonia. *In*:

RODGERS J. (ed.) *El Sistema Cambrico, su paleogeografia y el problema de base, 20th session, International Geological Congress, Mexico City*, 2, 97-126.

PANTIN, H.M. 1979. Interaction between velocity and effective density in turbidity flow with criteria for autosuspension. *Marine Geology*, 31, 59-99.

PATCHETT, P.J., GALE, N.H., GOODWIN, R. & HUMM, M.J. 1980. Rb-Sr whole-rock isochron ages of late Precambrian to late Cambrian igneous rocks from southern Britain. *Journal of the Geological Society, London*, 137, 649-56.

PATCHETT, P.J. & JOCELYN, J. 1979. U-Pb zircon ages for late Precambrian igneous rocks in South Wales. *Journal of the Geological Society, London*, 136, 13-19.

PETTIJOHN, F.J. 1957. *Sedimentary rocks*. Harper & bros., New York.

PETTIJOHN, F.J., POTTER, P.E. & SIEVER, R. 1972. *Sand and Sandstone*. Springer Verlag, New York.

PHILLIPS, STILLMAN C.J & MURPHY, 1976. A Caledonian plate tectonic model. *Journal of the Geological Society, London*, 132, 579-609.

PICKERING, K.T. 1980. *A Precambrian submarine fan and upper basin-slope succession in the Barents Sea Group, Finnmark, N. Norway*. D.Phil. thesis, University of Oxford.

PICKERING, K.T. & HISCOTT, R.N. 1985. Contained (reflected) turbidity currents from the Middle Ordovician Cloridorme Formation, Quebec, Canada: an alternative to the antidune hypothesis. *Sedimentology*, 32, 373-94.

PICKERING, K.T., STOW, D.A.V., WATSON, M.P. & HISCOTT, R.N. 1986. Deep water facies, processes and models: a review and classification scheme for modern and ancient sediments. *Earth Science Reviews*, 23, 75-174.

PIPER, D.J.W. 1970. A Silurian deep sea fan deposit in western Ireland and its bearing on the nature of turbidity currents. *Journal of Geology*, 78, 509-22.

PIPER, D.J.W. 1978. Turbidite muds and silts on deep-sea fans and abyssal plains. *In*: STANLEY, D.J. & KELLING, G. (eds.) *Sedimentation in submarine canyons, fans and trenches*. Dowden, Hutchinson & Ross, Stroudsburg, 163-76.

PIPER, J.D.A. 1982. The Precambrian palaeomagnetic record; the case for the Proterozoic supercontinent. *Earth and Planetary Science Letters*, 59, 61-89.

POTTER, T.L. 1974. *The stratigraphic palynology of some Cambrian successions in North Wales, England and North-West Spain*. Ph.D thesis, University of Sheffield.

POWELL, D.W. 1955. Gravity and magnetic anomalies in North Wales. *Quarterly Journal of the Geological Society, London*, 111, 375-97.

RAISON, C.A. 1891. On the lower limit of the Cambrian Series in North-West Caernarvonshire. *Quarterly Journal of the Geological Society, London*, 47, 329-42.

RAMSAY, A.C. 1866 & 1881. The geology of North Wales. *Memoir of the Geological Survey of Great Britain*, III.

RAST, N. 1962. Textural evidence for the origin of ignimbrites. *Liverpool & Manchester Geological Journal*, 3, 97-108.

RAST, N., O'BRIEN, B.H. & WARDLE, R.J. 1976. Relationships between Precambrian and Lower Palaeozoic rocks of the "Avalon Platform" in New Brunswick, the NE. Appalachians and the British Isles. *Tectonophysics*, 30, 315-338.

REEDMAN, A.J., LEVERIDGE, B.E. & EVANS, R.B. 1984. The Arfon Group ("Arvonian") of North Wales. *Proceedings of the Geologists' Association, London*, 95, 313-21.

REEDMAN, A.J., WEBB, B.C., ADDISON, R., LYNAS, B.D.T., LEVERIDGE, B.E. & HOWELLS, M.F. 1983. The Cambrian-Ordovician boundary between Aber and Betws Garmon, Gwynedd, North Wales. *Report of the Institute of Geological Sciences, London*, 83/1, 7-10.

RICE, R. & SHARP, G. 1976. Copper mineralisation in the forest of Coed-y-Brenin, Wales. *Transactions of the Institution of Mineralogists and Metallurgists*, Section B, 85, 1-13.

RICCI LUCCHI, F. 1984. *Fan sedimentation with emphasis on the North Sea Tertiary*. Joint Association for Petroleum Exploration Courses, Course notes, 29.

ROBERTS, B. 1979. *The geology of Snowdonia & Llyn: an outline and field guide*. Adam Hilger, Bristol.

ROBERTS, B. & MERRIMAN, R.J. 1985. The distinction between

Caledonian burial and regional metamorphism in metapelites from North Wales, an analysis of isocryst patterns. *Journal of the Geological Society, London*, 142, 615-25.

ROCHELEAU, M. & LAJOIE, J. 1974. Sedimentary structures in resedimented conglomerates of the Cambrian flysch, L'Islet, Quebec Appalachians. *Journal of Sedimentary Petrology*, 44, 826-36.

ROY, S. 1976. Ancient manganese deposits. *In*: WOLF, K.H. (ed.), *Handbook of Stratabound and Stratiform Ore Deposits*. 7, Elsevier, Amsterdam, 395-476.

RUSHTON, A.W.A. 1966. The Cambrian trilobites from the Purley Shales of Warwickshire. *Palaeontological Society* (Monograph), 120,

RUSHTON, A.W.A. 1974. The Cambrian of England and Wales. *In*: HOLLAND, C.H. (ed.) *Cambrian of the British Isles, Norden and Spitsbergen. (Lower Palaeozoic Rocks of the World, volume 2.)* John Wiley, London, 43-121.

RUSHTON, A.W.A. 1978. Description of the macrofossils from the Dolwen Formation. *In*: ALLEN, P.M & JACKSON, A.A. Bryn-teg Borehole, North Wales. *Bulletin of the Geological Survey of Great Britain*, 61, 46-48.

SCHUMM, S.A. 1960. The effect of sediment type on the shape and stratification of some modern fluvial deposits. *American Journal of Science*, 258, 177-84.

SEDGWICK, A. 1843. Outline of the geology of North Wales. *Proceedings of the Geological Society, London*, 4, 212-24.

SELLEY, R.C. 1985. *Ancient sedimentary environments*. Chapman & Hall, London.

SEILACHER, A. 1967. Bathymetry of trace fossils. *Marine Geology*, 5, 413-28.

SHACKLETON, R.M. 1953. The structural evolution of North Wales. *Liverpool & Manchester Geological Journal*, 1, 261-97.

SHACKLETON, R.M. 1956. Notes on the structure and relations of the Precambrian and Ordovician rocks of south-western Lleyn (Caernarvonshire). *Liverpool & Manchester Geological*

Journal, 1, 400–9.

SHACKLETON, R.M. 1975. Precambrian rocks of Wales. *In*: HARRIS, A.L., SHACKLETON, R.M., WATSON, J., DOWNIE, C., HARLAND, W.B. & MOORBATH, S. (eds.) *A correlation of the Precambrian rocks in the British Isles*. Geological Society, London, Special Report, 6, 76–82.

SHANNON, P.M. 1978. The stratigraphy and sedimentology of the Lower Palaeozoic rocks of South-East County Wexford. *Proceedings of the Royal Irish Academy*, Section B. 78, 247–65.

SHANNON, P.M. 1980. Depositional mechanisms and environments in the Lower Palaeozoic clastic sediments of SE. Ireland. *Journal of Earth Science, Royal Dublin Society*, 2, 97–114.

SHEPARD, F.P. & MARSHALL, N.F. 1978. Currents in submarine canyons and other sea valleys. *In*: STANLEY, D.J. & KELLING, G. (eds.) *Sedimentation in submarine canyons, fans and trenches*. Dowden, Hutchinson & Ross, Stroudsburg, 3–14.

SMITH, D.G. 1977. Lower Cambrian palymorphs from Howth, Co. Dublin. *Geological Journal*, 12, 159–68.

SMITH, N.D. 1971. Pseudo-planar stratification produced by very low amplitude sand waves. *Journal of Sedimentary Petrology*, 41, 69–73.

SMITH, R.L. & BAILEY, R.A. 1968. Resurgent cauldrons. *In*: COATES, R.R., HAYES, R.L. & ANDERSON, C.A. (eds.) *Studies in Volcanology*. Geological Society of America Memoir, 116, 613–62.

SOPER, N.J. & HUTTON, D.H.W. 1984. Late Caledonian sinistral displacement in Britain: implications for a three-plate collision model. *Tectonics*, 3, 781–94.

STAUFFER, P.H. 1967. Grain-flow deposits and their implications, Santa Ynez Mts., California. *Journal of Sedimentary Petrology*, 37, 487–508.

STILLMAN, C.J. 1986. A comparison of the Lower Palaeozoic volcanic rocks on either side of the Caledonian suture in the British Isles. *In*: FETTES, D.J. & HARRIS, A.L. (eds.) *Synthesis of the Caledonian rocks of Britain*.

Reidel, Dordrecht, 187-205.

STOW, D.A.V. 1982. Bottom currents and contourites in the North Atlantic. *Actes Colloque International C.N.R.S. Bordeaux. Bull. Inst. Geol. Bassin d'Aquitaine, Bordeaux*, 151-66.

STOW, D.A.V. 1985. Deep sea clastics: where are we now and where are we going? *In*: BRENCHLEY, P.J & WILLIAMS, B.P.J. (eds.) *Sedimentology: recent developments and applied aspects*. Special publication of the Geological Society, London, No. 18, 67-93.

STOW, D.A.V. & HOLBROOK, J.A. 1984. North Atlantic contourites: an overview. *In*: STOW, D.A.V. & PIPER, D.J.W. (eds.) *Fine grained sediments: deep water processes and facies*. Special publication of the Geological Society, London, No. 15, 245-56.

STOW, D.A.V. & LOVELL. J.P.B. 1979. Contourites: their recognition in modern and ancient sediments. *Earth Science Reviews*, 14, 251-91.

STOW, D.A.V. & SHANMUGAM, G. 1980. Sequence of structures in fine grained turbidites: comparison of recent and ancient flysch sediments. *Sedimentary Geology*, 25, 23-42.

STRONG, P.G. & WALKER, R.G. 1981. Deposition on the Cambrian continental rise: the St. Roch Formation near St. Jean-Port-Joli, Quebec. *Canadian Journal of Earth Science*, 18, 1320-35.

STUBBLEFIELD, C.J. 1956. Cambrian Palaeogeography in Britain. *Congreso Geologico International, XX Sesion, Mexico.*

SURLYK, F. 1978. Submarine fan sedimentation along fault scarps on tilted fault blocks (Jurassic-Cretaceous boundary, E. Greenland). *Gronlands Geologiske Undersogelse Bulletin*, 128.

TAYLOR, H.P. 1979. Oxygen and hydrogen isotope relationships in hydrothermal mineral deposits. *In*: BARNES, H.L. (ed.) *Geochemistry of Hydrothermal Ore deposits*. John Wiley, New York, 236-77.

THORPE, R.S. 1972. Ocean floor basalt affinity of Precambrian glaucophane schist from Anglesey. *Nature*, 240, 164-66.

THORPE, R.S. 1974. Aspects of magmatism and plate tectonics in the Precambrian of England and Wales. *Geological Journal*, 9, 115-36.

THORPE, R.S., BECKINSALE, R.D., PATCHETT, P.J., PIPER, J.D.A. DAVIES, G.R. & EVANS, J.A. 1984. Crustal growth and late Precambrian-early Palaeozoic tectonic evolution of England and Wales. *Journal of the Geological Society, London*, 141, 521-36.

TURNER, P. 1978. Diagenetic origin of Cambrian marine red beds: Caerfai Bay shales, Dyfed, Wales. *Sedimentary Geology*, 24, 269-81.

UNRUG, R. 1963. Istebna beds- a fluxoturbidity formation in the Carpathian flysch. *Annales de la Societe Geologique de Pologne*, 33, 49-92.

VALENTINE, P.C., COOPER, R.A. & UZMANN, J.R. 1984. Submarine sand dunes and sedimentary environments in Oceanographer canyon. *Journal of Sedimentary Petrology*, 54, 704-15.

VAN LUNSEN, H.A. & MAX, M.D. 1975. The geology of Howth and Ireland's Eye, Co. Dublin. *Geological Journal*, 10, 35-58.

WALKER, R.G. 1965. The origin and significance of the internal sedimentary structures of turbidites. *Proceedings of the Yorkshire Geological Society*, 35, 1-32.

WALKER, R.G. 1967. Turbidite sedimentary structures and their relationship to proximal and distal depositional environments. *Journal of Sedimentary Petrology*, 37, 25-43.

WALKER, R.G. 1970. Review of the geometry and facies organisation of turbidites and turbidite-bearing basins. *In*: LAJOIE, J. (ed.) *Flysch Sedimentology in North America*. Geological Association of Canada, Special Paper, 7, 219-51.

WALKER, R.G. 1978. Deep water sandstone facies and ancient submarine fans: models for exploration for stratigraphic traps. *American Association of Petroleum Geologists, Bulletin*, 62, 932-66.

WATSON, J. 1984. The ending of the Caledonian orogeny in

Scotland. *Journal of the Geological Society, London,* 141, 193-214.

WEBB, B.C. 1983. Early Caledonian structures in the Cambrian slate belt, Gwynedd, North Wales. *Report of the Institute of Geological Sciences, London,* 83/1, 1-6.

WELLS, A.K. 1925. The geology of the Rhobell Fawr district (Merioneth). *Quarterly Journal of the Geological Society, London,* 81, 463-538.

WILLIAMS, D. 1923. *On the Geology of a part of Northern Snowdonia.* MSc. thesis, University of Liverpool.

WILLIAMS, D. 1930. The geology of the country between Nant Peris and Nant Ffrancon (Snowdonia). *Quarterly Journal of the Geological Society, London,* 86, 191-233.

WILSON, I.G. 1971. Desert sandflow basins and a model for the development of ergs. *Geographical Journal,* 137, 180-199.

WILSON, J.T. 1966. Did the Atlantic close and then re-open? *Nature,* 211, 676-81.

WILSON, M.D. & SEDEORA, S.S. 1979. An improved thin section stain for potash feldspar. *Journal of Sedimentary Petrology,* 49, 637-38.

WINN, R.D. & DOTT, R.H. 1977. Large-scale traction-produced structures in deep-water fan-channel conglomerates in southern Chile. *Geology,* 5, 41-44.

WINN, R.D. & DOTT, R.H. 1979. Deep-water fan-channel conglomerates of late Cretaceous age, southern Chile. *Sedimentology,* 26, 203-228.

WOOD, D.S. 1969. The base and correlation of the Cambrian rocks of North Wales. *In*: WOOD, A. (ed.) *The Precambrian and Lower Palaeozoic Rocks of North Wales.* University of Wales Press, 47-66.

WOOD, D.S. 1973. Patterns and magnitudes of natural strain in rocks. *Philosophical Transactions of the Royal Society of London,* Series A, 274, 373-82.

WOOD, D.S. 1974. Ophiolites, melanges, blueschists and ignimbrites: early Caledonian subduction in Wales? *In*: DOTT, R.H. & SHAVER, R.H. (eds.) *Modern and Ancient Geosynclinal Sedimentation.* Special Publication of the

Society of Economic Paleontologists & Mineralogists, Tulsa, No. 19, 334-44.

WOODCOCK, N.H. 1984a. The Pontesford Lineament, Welsh Borderland. *Journal of the Geological Society, London*, 141, 1001-14.

WOODCOCK, N.H. 1984b. Early Palaeozoic sedimentation and tectonics in Wales. *Proceedings of the Geologists' Association, London*, 95, 323-35.

WOODLAND, A.W. 1938a. Petrographical studies in the Harlech Grits Series, Merionethshire. I.: metamorphic changes in the mudstones of the Manganese Shale Group. *Geological Magazine*, 75, 366-82.

WOODLAND, A.W. 1938b. Petrological studies in the Harlech Grit Series of Merionethshire. II.: the petrography and petrology of some of the grits. *Geological Magazine*, 75, 440-54.

WOODLAND, A.W. 1938c. Some petrological studies in the Harlech Grit Series of Merionethshire. III.: the development of pyrite in the grits and mudstones. *Geological Magazine*, 75, 529-39.

WOODLAND, A.W. 1939. Petrography and petrology of the Lower Cambrian manganese ore of W. Merionethshire. *Quarterly Journal of the Geological Society, London*, 95, 1-35.

WOODLAND, A.W. 1956. The manganese deposits of Great Britain. *In: Symposium del Manganeso XX Congreso Geologico Internacional*, 199-218.

WOODLAND, A.W. 1971. (ed.) The Llanbedr (Mochras Farm) Borehole. *Report of the Institute of Geological Sciences*, 71/18.

WOODWARD, H.B. 1888. On the discovery of trilobites in the Upper Green (Cambrian) Slates of the Penrhyn Quarries, Bethesda. *Quarterly Journal of the Geological Society, London*, 44, 74-8.

WRIGHT, A.E. 1977. The evolution of the British Isles in the late Precambrian. *Estudios Geologicos*, 33, 303-313.

ZIEGLER, A.M. & MCKERROW, W.S. 1975. Silurian Marine Red Beds. *American Journal of Science, 275, 31-56.*

APPENDICES

APPENDIX 1 : Carbonate Carbon and Oxygen Stable Isotope Analysis.

Sample Preparation.

Carbon and oxygen stable isotopic ratios of rhodochrosite samples were determined by reaction with 100% orthophosphoric acid, under vacuum, at 25°C using a method similar to that described by McCrea (1950). Evolved carbon dioxide was cleaned of water and phosphoric acid vapour in a cryogenic acetone trap (-87°C) and collected in evacuated collection vessels at liquid nitrogen temperatures (c. -190°C). Non condensable phases (O_2, N_2) were removed under vacuum prior to closure of the sample vessel.

Analytical Procedures.

Samples were analysed along with simultaneously prepared standards against an internal reference gas on a Sira 9 triple collecting mass spectrometer. Raw gas values were corrected for instrumental, isobaric and ^{17}O effects following the procedures of Craig (1957) and Deines (1970). All data were corrected for the temperature controlled reaction:

orthophosphoric acid + rhodochrosite (manganese carbonate) -> carbon dioxide using a fractionation factor (α) of 1.01012 (Friedman & O'Neill 1977). Isotopic ratios are reported in parts per mil difference between the sample and the PDB standard in conventional notation:

$$\text{e.g del } ^{13}C = \frac{(^{13}C/^{12}C_{sample} - {}^{13}C/^{12}C_{std})}{(^{13}C/^{12}C_{std})} \times 1000$$

The reproducibility of the analytical technique was estimated on the basis of repeated analysis of a single whole rock sample of unknown composition. Average reproducibility of the technique was determined as better than 0.03 per mil and 0.08 per mil for del ^{13}C and del ^{18}O respectively.

TABLE OF ISOTOPE RESULTS.

Sample Type	Del 13C	Del 18O
Whole Rock	-12.930	-17.320
Whole Rock	-12.956	-17.484
Whole Rock	-12.962	-17.532
Whole Rock	-12.893	-17.386
Whole Rock	-12.924	-17.441
Red Band	-14.669	-15.467
Light Band	-12.546	-19.238
Red Band	-13.339	-17.016
Light Band	-12.343	-18.969
Red Band	-14.035	-16.389

APPENDIX 2. POINT COUNT DATA.

Each thin section was point counted for 500 points, the number of counts for each category was recorded.

KEY

Formation.

DwG	Dolwen Formation	Harlech Dome
RnG	Rhinog Formation	Harlech Dome
HfF	Hafotty Formation	Harlech Dome
BmG	Barmouth Formation	Harlech Dome
GmF	Gamlan Formation	Harlech Dome
CCG	Cefn Coch Grit	Harlech Dome
HMG	Hell's Mouth Grits	St Tudwal's Peninsula
MnB	Mulfran Beds	St Tudwal's Peninsula
CnG	Cilan Grits	St Tudwal's Peninsula
BrG	Bronllwyd Grit Formation	Arfon

Composition.

Q_m	Monocrystalline Quartz
Q_p	Polycrystalline Quartz
Total Q	Total Quartz
Total F	Total Feldspar
L_v	Volcanic Lithic Fragments
L_p	Plutonic Lithic Fragments
L_s	Sedimentary and Metamorphic Lithic Fragments
Total L	Total Lithic Fragments
Matrix	Matrix

Rock Name. (Pettijohn *et al.* 1972)

FG	Feldspathic Greywacke
LG	Lithic Greywacke
QW	Quartz Wacke
QA	Quartz Arenite
LA	Lithic Arenite
SLA	Sublitharenite

SA	Subarenite

Grain Size.

S	Sandstone
C	Conglomerate
vf	very fine
f	fine
m	medium
c	coarse
vc	very coarse

Sample Number	Formation	Qm	Qp	Total Q	Total F	Lv	Lp	Ls	Total L	Matrix	Rock Name	Grain Size
69021	DwG	68	228	296	39	9	71	63	143	22	LA	GC
69022	DwG	337	62	399	20	1	15	3	19	62	QA	fS
69023	DwG	247	80	327	65	1	20	16	37	71	SA	fS
69024	DwG	217	156	373	0	0	1	16	17	110	QW	mS
65911	RnG	259	78	337	29	1	16	3	20	114	FG	mS
65912	RnG	263	110	373	32	14	43	1	58	37	SLA	cS
65913	RnG	263	54	317	29	8	6	2	16	138	FG	mS
65915	RnG	182	118	300	34	53	7	18	78	88	LG	cS
65916	RnG	235	19	254	13	2	0	1	3	230	QG	vfS
65917	RnG	233	49	282	18	4	0	2	6	194	FG	mS
65918	RnG	171	137	308	21	2	2	107	111	60	LA	cS
65919	RnG	191	60	251	12	27	8	147	182	55	LA	PC
65920	RnG	152	35	187	54	124	2	2	128	131	LG	cS
65921	RnG	224	44	268	38	67	3	0	70	124	LG	mS
65922	RnG	200	89	289	12	113	6	7	126	73	LA	vcS
65923	RnG	174	32	206	20	139	11	2	152	122	LG	cS
65924	RnG	150	52	202	27	170	10	29	209	62	LA	GC
65938	RnG	170	33	203	37	1	8	8	17	243	FG	cS
65939	RnG	239	130	369	31	10	18	12	40	60	SLA	cS
65940	RnG	137	33	170	19	3	1	4	8	303	FG	mS
65941	RnG	199	82	281	35	8	8	5	21	163	FG	cS
65976	RnG	229	46	275	49	8	9	0	17	159	FG	cS
67786	RnG	163	23	186	51	76	1	6	83	180	LG	mS
67787	RnG	207	23	230	51	55	2	0	57	162	LG	mS
67788	RnG	205	59	264	38	80	11	23	114	84	LG	mS
67789	RnG	190	46	236	32	97	9	9	115	117	LG	mS
67790	RnG	140	114	254	28	84	26	31	141	77	LG	vcS
67791	RnG	264	79	343	27	49	26	19	94	36	SLA	vcS
67792	RnG	235	66	301	32	109	12	19	140	27	LA	vcS
67793	RnG	178	95	273	25	127	18	14	159	43	LA	GC
67794	RnG	248	84	332	49	6	12	2	20	99	FG	mS
67795	RnG	129	100	229	21	168	18	10	196	54	LA	GC
67796	RnG	241	37	278	15	43	0	3	46	161	LG	mS
67797	RnG	236	52	288	2	9	0	102	111	99	LG	mS
67800	RnG	266	92	358	40	3	17	0	20	82	FG	cS
67801	RnG	281	85	366	42	11	41	12	64	28	SLA	cS
67802	RnG	262	15	277	24	2	0	3	5	194	FG	mS
69011	RnG	251	58	309	30	100	14	10	124	37	LA	vcS
69012	RnG	56	191	247	25	159	10	14	183	45	LA	PC
69013	RnG	101	104	205	34	149	18	28	195	66	LA	PC
65942	HfF	245	32	277	30	3	6	2	11	182	FG	cS
67806	HfF	209	27	236	48	2	1	7	10	206	FG	cS
65926	BmG	217	48	265	22	0	36	7	43	170	LG	mS
65927	BmG	228	56	284	29	0	21	19	40	147	LG	mS
65928	BmG	134	137	271	22	6	57	4	67	140	LG	GC
65930	BmG	255	85	340	46	0	8	1	9	105	FG	mS
65931	BmG	257	94	351	30	1	20	1	22	97	FG	mS
65932	BmG	137	58	195	24	5	108	34	147	134	LG	GC
65933	BmG	80	52	132	37	0	41	254	295	36	SLA	PC
65935	BmG	185	86	271	36	3	52	1	56	137	LG	mS
65936	BmG	226	48	274	59	0	50	3	53	114	FG	mS
65937	BmG	253	21	274	53	3	1	3	7	166	FG	mS
65948	BmG	214	135	349	45	0	23	3	26	80	FG	cS
65973	BmG	168	73	241	57	11	93	24	128	74	LA	GC
65974	BmG	184	127	311	40	17	25	0	42	107	LG	cS
65975	BmG	183	93	276	67	0	45	2	47	110	FG	cS
65977	BmG	178	75	253	49	0	55	0	55	143	LG	vcS

Sample Number	Formation	Qm	Qp	Total Q	Total F	Lv	Lp	Ls	Total L	Matrix	Rock Name	Grain Size
65946	GmF	140	52	192	20	2	7	1	10	278	FG	cS
65947	GmF	199	29	228	37	1	8	1	10	225	FG	mS
67808	GmF	168	115	283	12	0	54	0	54	151	LG	vcS
67809	GmF	120	61	181	35	48	62	28	138	146	LG	vcS
67810	GmF	153	95	248	39	0	84	39	123	90	LG	vcS
69016	GmF	191	71	262	51	4	58	48	110	77	LG	cS
69017a	GmF	219	14	233	35	0	0	6	6	226	FG	mS
69017b	GmF	245	14	259	28	0	2	3	5	208	FG	mS
69018a	GmF	235	58	293	56	0	11	3	14	137	FG	mS
69018b	GmF	259	52	311	48	0	8	5	13	128	FG	mS
69019a	GmF	245	104	349	50	4	2	10	16	85	FG	cS
69019b	GmF	206	96	302	63	0	31	3	34	101	FG	vcS
69020a	GmF	254	101	355	41	1	4	6	11	93	FG	cS
69020b	GmF	237	89	326	50	0	5	4	9	115	FG	cS
69014	CCG	333	67	400	2	0	3	6	9	89	QW	cS
69015	CCG	350	54	404	2	0	0	14	14	80	QW	cS
64559	HMG	249	38	287	40	12	10	0	22	151	FG	mS
64560	HMG	233	68	301	29	31	41	14	86	84	LG	cS
64561	HMG	311	41	352	19	13	7	1	21	108	LG	mS
64563	HMG	232	50	282	38	35	1	9	45	135	LG	mS
64564	HMG	314	72	386	12	13	16	0	29	73	SLA	mS
64567	HMG	259	40	299	40	21	10	0	31	130	FG	mS
64568	HMG	241	74	315	23	39	13	0	52	110	LG	cS
64569	HMG	250	79	329	31	44	10	0	54	86	LG	vcS
64570	HMG	244	60	304	33	37	25	9	71	92	LG	cS
64572	HMG	316	42	358	30	19	2	0	21	91	FG	cS
64573	HMG	299	51	350	30	24	10	0	34	86	LG	fS
64576	HMG	304	60	364	47	14	14	5	33	56	SA	cS
64577	HMG	242	28	270	32	13	6	0	19	179	FG	mS
64578	HMG	173	12	185	31	8	2	0	10	274	FG	fS
64579	HMG	240	5	245	38	5	4	2	11	206	FG	fS
64580	HMG	295	44	339	28	25	4	1	30	103	LG	cS
67048	HMG	246	65	311	22	16	35	0	51	116	LG	cS
64583	MnB	311	8	319	35	9	1	3	13	133	FG	mS
64588	CnG	214	49	263	21	8	16	0	24	192	LG	cS
64590	CnG	172	24	196	22	1	6	1	8	274	FG	mS
64591	CnG	239	77	316	51	3	47	1	51	82	LG	cS
64592	CnG	221	33	254	49	4	28	5	37	160	FG	mS
64593	CnG	269	32	301	44	7	6	1	14	141	FG	mS
64597	CnG	233	35	268	48	3	26	4	33	151	FG	mS
65962	CnG	215	52	267	31	3	37	91	131	71	LA	cS
65967	CnG	300	31	331	52	0	11	3	14	103	FG	cS
65968	CnG	282	64	346	42	1	1	12	14	98	FG	mS
69009a	CnG	221	99	320	41	9	27	33	69	70	SLA	cS
69009b	CnG	305	39	344	44	11	5	12	28	84	FG	mS
69010	CnG	290	28	318	35	1	3	0	4	143	FG	fS
69025	BrG	253	68	321	37	2	37	0	39	103	LG	cS
69026	BrG	269	55	324	37	3	13	1	17	122	FG	mS
69027	BrG	273	48	321	38	3	8	0	11	130	FG	cS
69028	BrG	191	88	279	55	3	59	0	62	104	LG	cS
69029	BrG	226	73	299	48	6	31	0	37	116	FG	cS
69030a	BrG	211	25	236	35	1	94	0	95	134	LG	GC
69030b	BrG	233	15	248	45	0	2	0	2	205	FG	fS
69031	BrG	137	118	255	32	2	92	0	94	119	LG	GC
69032	BrG	246	50	296	30	1	6	0	7	167	FG	cS
69033	BrG	190	68	258	54	2	31	0	33	155	FG	cS
69034	BrG	215	71	286	49	4	37	1	42	123	FG	cS