THE STRATIGRAPHY OF THE FISHGUARD VOLCANIC COMPLEX AND ASSOCIATED SEDIMENTS, SW DYFED, WALES.

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Doctor of Philosophy

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SUMMARY

This thesis describes the geology of a Lower Palaeozoic terrain, situated west of the town of Fishguard, SW Dyfed, Wales. The area is dominated by the Fishguard Volcanic Complex (Upper Llanvirn), and sediments that range in age from the Middle Cambrian to the Lower Llandeilo. The successions represent an insight into sedimentation and volcanism for *c*. 100 Ma. along the south-western margin of the Lower Palaeozoic Welsh Basin.

The stratigraphy of the sedimentary sequence has been completely revised and the existing volcanostratigraphy modified. The observed complexity of the stratigraphy is primarily the consequence of Caledonide deformation which resulted in large scale repetition. Fold–thrust tectonics dominates the structural style of the area. Caledonide trending (NE-SW) cross-faults complicate preexisting structures.

Middle Cambrian (?) sedimentation is documented by shallow marine clastics and red shales deposited within tidal – subtidal environments. Upper Cambrian sedimentation was dominated by shallow marine 'storm' and 'fair weather' sedimentation within a muddy shelf environment. Shallow marine conglomerates and heterolithic intertidal siliciclastics mark the onset of Ordovician sedimentation during the lower Arenig transgression. Mid-Arenig sediments reflect deposits influenced by storm, fair-weather and wave related processes in various shallow marine environments, including; shoreface, inner shelf, shoaling bar, and deltaic. Graptolitic marine shales were deposited from the upper mid-Arenig through to the Lower Llandeilo; during which time sediments accumulated by pelagic processes and fine grained turbidites. The varied nature of sedimentation reflect both localised change within the depositional system and the influence of larger regional eustatic events.

Ordovician subaqueous volcanic activity produced thick accumulations of lavas, pyroclastics, hydroclastics, and hyaloclastics. The majority of volcanism was effusive in nature, erupted below the Pressure Compensation Level. Basaltic volcanism was characterised by pillowed lavas and tube networks, whilst sheet-flow lavas, pillow breccias and minor hyaloclastites developed locally. Silicic volcanism was dominated by rhyolitic clastics of various affinities, although coherent silicic obsidian lavas, sheet-flow lavas and pyroclastics developed. Hypabyssal intrusives of variable composition and habit occur throughout the volcanic successions.

Low–grade regional metamorphism has variably affected the area, conditions of the prehnite–pumpellyite and greenschist facies having been attained. Numerous secondary phases developed in response to the conditions imposed, which collectively indicate that P–T conditions were of low-pressure facies series in the range P= 1.2–2.0 kbars and T= 230–350°C, under an elevated geothermal gradient of 40–45°C km^{-1}. Polymineralic cataclastites associated with Caledonide deformation indicate that tectonism and metamorphism were in part contemporaneous.

Key Words: Subaqueous Volcanism   Low-grade Metamorphism   Sedimentology   SW Dyfed   Welsh Basin   Lower Palaeozoic

To Sue.

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CONTENTS

CHAPTER FOUR

SEDIMENTARY STRATIGRAPHY AND SEDIMENTATION

FIGURES, TABLES, PLATES, AND MAPS.

CHAPTER ONE.

CHAPTER TWO

CHAPTER THREE

CHAPTER FOUR

CHAPTER FIVE

CHAPTER SIX

CHAPTER SEVEN

ENCLOSURES (in back pocket of thesis)

Duplicate stratigraphic tables and summary diagrams

TEXT.

CHAPTER ONE
REGIONAL SETTING, LOCATION, AIMS, PREVIOUS RESEARCH

1.1. REGIONAL SETTING (The Lower Palaeozoic Welsh Basin)

The Lower Palaeozoic Welsh Basin (Fig 1.1.), an ensialic back-arc/marginal basin situated SE of the Iapetus Ocean, was founded during late Precambrian/early Cambrian times within the mobile Precambrian micro-continent of Avalonia; one of three cratonic/fold-belt terrains (Fig. 1.1.) which presently constitute the Appalachian – Southern European Caledonides (Kennedy 1979, Kokelaar *et al.* 1984a, Mckerrow & Cocks 1986, André *et al.* 1986). From inception the Basin was to exist as a sedimentary prism until late Silurian/early Devonian times, where upon final closure of the Iapetus Ocean resulted in collision of the northerly Lauratian Landmass and Southern European Caledonides culminating in the Caledonian Orogeny and uplift of the accumulated basin-fill.

Throughout much of its Lower Palaeozoic history, the Welsh Basin was broadly defined by deep-seated NE–SW trending pre-orogenic lineaments; the NW margin being bounded by the Menai Strait Fault System and periodically emergent Anglesey Horst, whilst to the SE the basin margin was defined by the Pontesford-Lindley – Church Stretton lineament and passive Midland Platform Craton (Fig. 1.1). In character with modern marginal basins, such lineaments and intra-basinal structural systems (i.e. Bala Fault System; Fitches & Cambell 1987) controlled depositional patterns and sites of volcanism.

Much of early Lower Palaeozoic (Cambrian/mid-Ordovician) sedimentation was characterised by shallow marine clastics punctuated by prolonged periods of mudstone deposition; the latter commonly reflecting global eustatic events and localised 'adjustment' to extension and intra-basinal rifting associated with the major phases of volcanism. Thick black marine shales dominate much of basins mid-Ordovician to late-Silurian history, reflecting a deep and possibly restricted basin environment, although it is increasingly advocated that abyssal depths are inappropriate.

Igneous activity within the Welsh Basin is principally recorded from the Tremadoc through to Caradoc times. Early activity (Tremadoc-Arenig) was characterised by basalt-andesite-dacite-rhyodacite magmas with island-arc affinities (i.e Rhobell Volcanic Group, Trefgarne Volcanic Group) followed by the widespread local development of bimodal (acid-basic) marginal basin tholeiities (e.g. Fishguard Volcanic Complex); during which time crustal tension was prevalent although no oceanic lithosphere developed (Bevins *et al.* 1984, Kokelaar *et al.* 1984a).

Burial and deformation resulted in low-grades of metamorphism affecting the volcanosedimentary basin-fill. Conditions varied from the zeolite facies to the prehnite-pumpellyite facies in marginal shelf areas, whilst central parts of the basin reached the lower greenschist facies (Bevins & Rowbotham 1983); nowhere did conditions approach the amphibolite facies (i.e. T> 380°C).

Despite extensive recent research throughout the Welsh Basin (Fitches & Woodcock *conference report* 1987; Fitches & Woodcock *eds*. 1987) much of the southern 'marginal/shelf' terrain of SW Dyfed remains undescribed. In this thesis a small area of this terrain is considered (Fig. 1.1.), and which is sufficiently heterogeneous so as to allow varying aspects of 'Welsh Basin' geology to investigated.

Figure 1.1. Geological map of the Lower Palaeozoic Welsh Basin, showing the distribution of major lineaments, stratigraphic successions, and Ordovician volcanic centres (*after*, Kokelaar *et al*. 1984a)

1.2. LOCATION OF STUDY AREA

This study is concerned with an investigation of the Lower Palaeozoic geology, west of the town of Fishguard, SW Dyfed, Wales (Fig. 1.2., Maps 1-6); the following work also considers the relevant geological history of surrounding Lower Palaeozoic strata.

The principle study area covers approximately 90 sq.km., which includes the coastline between Fishguard and Porth-gain, whilst inland it covers the Pen Caer peninsular, Dyffryn, Manorowen, St. Nicholas, and Granston districts (Fig. 1.2.). The topography is common to much of SW Wales, being dominated by a series of remnant pediplained platforms and upstanding ridges; features attributed to pediplaination and deglaciation ice thinning related to the late Devensian Ice Sheet *c.* 18000 - 15000 y.a. (*c.f.* Bowen 1982). The effects of glaciation has been to produce steep- cliffed coastline, indentation of which frequently reflects lithological or structural change. Inland, thick glacial cover results in almost complete absence of exposure, outcrop being restricted to a few isolated localities, road cuts, and infrequent quarries, effectively precluding detailed study. Uncharacteristically, feature mapping was found to be of limited use, however, new generalised geological maps of the Fishguard – Porth-gain district are presented (Maps 1-4), forming the basis of this study.

Figure 1.2. Location map of study area in SW Dyfed, showing the position of the principle settlements commonly referred to in the following text.

1.3. PROJECT AIMS & GENERAL GEOLOGY

Project aims

The geology of the Fishguard – Porth-gain district is comprised of igneous and sedimentary rocks believed to range in age from the Middle Cambrian to the Upper Llandeilo, all of which have been subjected to low-grades of metamorphism and deformed during the Caledonian Orogeny.

Spatially the study area is dominated by bimodal (acid-basic) volcanics and contemporaneous intrusives which collectively form part of the Fishguard Volcanic Complex (Fig. 1.3.). The original objective of this project was to investigate the development of the Volcanic Complex, west of Fishguard. However, it became apparent that whilst the Complex may require protracted study, the 'regional' geology between Fishguard and Porth-gain was in part poorly understood. A multi-disciplinary study was therefore pursued in an attempt to establish conditions and environments of deposition, igneous processes, P-T conditions of metamorphism, and tectonic history within the confines of a 'working' lithostratigraphy. In short the project has no deliberate specialised aims, although emphasis is placed on the newly proposed stratigraphy which form a chronologic framework for the understanding of the area's early Lower Palaeozoic history.

General Geological Features

The Fishguard Volcanic Complex, a mid-Ordovician (Llanvirn) bimodal tholeiitic volcanic complex (Bevins 1982) represents the thickest development (>1500m) of igneous rock in S Wales, cropping out in a laterally continuously belt (>25 km) through the Prescelly – Fishguard region (Fig. 1.3.). For its most part the Complex developed subaqueously, the marine environment and bimodal magmas strongly influencing the style of eruptive mechanisms and nature of resultant deposits. Silicic volcanism was characterised by low lava/clast ratios with the development of an array of rhyolitic clastic facies rock types and subordinate coherent flows. Basaltic volcanism was effusive in nature developing extensive pillowed flow networks with only a minor hyaloclastics/hydroclastics component. Contemporaneous gabbros and tonalites (and finer grained equivalents) where emplaced at high-levels into the volcanic pile and surrounding sediments.

Lower Palaeozoic sediments believed to range in age from the Middle Cambrian to the Upper Llandeilo, occupy much of the coastal district. Conspicuous red-green shallow marine clastics of various affinity dominate the likely Middle Cambrian successions (Lach Dafad Formation, Fig. 1.3.). Upper Cambrian sediments (*Lingula* Group, Fig.1.3) are comprised of varying heterolithic sandstone-mudstone alternations exhibiting features and bed forms common to a siliciclastic shallow shelf setting. The onset of Ordovician sedimentation is

Figure 1.3. Geological sketch map of the Fishguard – Porth-gain district, showing the distribution of the major lithological divisions

recorded by sediments of Arenig age (Trwyn Llwyd Formation and Abercastle Formation, Fig. 1.3.), Tremadoc strata being absent. Lower Arenig sediments are comprised of a variety of shallow marine clastics and 'heterolithics', recording the development of a complex depositional system. Deposition of arenaceous material appears to have abated during the middle Arenig, from which time deposition was dominated by black marine mudstones within a 'stagnant' marine environment; conditions that were to persist until the late Upper Llanvirn (Arenig-Llanvirn Shales Fig. 1.3). A temporal influx of coarse terrigenous material is recorded at the end of the Llanvirn/Lower Llandeilo, where upon black marine mudstones where again deposited (Fishguard Bay Group, Fig 1.3.).

Low-grades of metamorphism have variably affected the area. Based on a consideration of Ca–Al secondary silicates from basic rock types within the Fishguard Volcanic Complex, conditions consistent with the prehnite-pumpellyite and greenschist facies prevailed. The generation of such conditions is believed to have arisen primarily as a consequence of burial related processes, although a wide range of factors appears to have been influential, including among others Caledonian deformation.

The broad structural framework of the area is defined by a disrupted anticline-syncline pair (Fig. 1.2.), the 'Llwanda Anticline' and 'Goodwick Syncline' (Cox 1930). However, the area has been subjected to complex deformation within which two discrete Caledonide phases can be identified. The first and most important, is seen as a possibly prolonged regional fold–thrust event responsible for large scale repetition in the volcano- sedimentary stratigraphy, the second phase is documented by NE–SW trending (Caledonide trend) faults. The latter generation of structures are geomorphically expressed, although, rarely do they have any major control on the regions geology other than complicating an already complex picture.

1.4. HISTORY OF RESEARCH

The Lower Palaeozoic geology of SW Dyfed has been a subject of interest since the beginning of the 18th century. Early studies established many fundamentals of the Cambrian and Ordovician Systems attracting the outstanding geologists of the day, including among others: de la Béche, Harkness and Hicks. At the turn of the century, A.H. Cox, O.T. Jones and H.H. Williams, helped lay the stratigraphic foundation which culminated in the geosycline model of Jones (1934) and the concept of a Lower Palaeozoic Welsh Basin.

Co-ordinated under the direction of Sir Henry de la Béche, the earliest work in the Fishguard – Porth-gain district is recorded on the first geological map of Pembrokeshire published in 1845 and subsequently revised in 1857 (1 inch geological map, Sheet 40, Old Series 1st Edit, rev 1857). On this map the area is variously defined as *b2 Llandeilo, feldspathic ashes and traps,* and *greenstones*; lithologies which refer in part to the Fishguard Volcanic Complex.

Early investigations of the Volcanic Complex are recorded in the accounts of Reed (1895), Elsden (1905) Cox & Jones (1913), and Cox (1930). The first detailed studies were made by Evans (1945) regarding the development of the Volcanic Complex in the Prescelly Hills, whilst Thomas & Thomas (1956) detailed the volcanics west of Fishguard; more specific studies include that of Evans (1954) who documents a welded ash-flow tuff east of Fishguard and Jones (1969) who discussed the vesicularity of pillow lavas.

More recently the principle aims of investigations have been to discuss operative volcanic processes (Bevins 1979; Bevins & Roach 1979a,b; Bevins & Roach 1982; Kokelaar *et al.* 1984a,b; Lowman 1977; Lowman & Bloxam 1981) and to model the Fishguard Volcanic Complex in the context of its geochemical signature and petrogenetic affinity with volcanic centres elsewhere in Wales (Lowman 1977, Bevins 1979, Bevins 1982, Bevins *et al.* 1984, Kokelaar *et al.* 1984a). Substantial parts of the above work is based on the doctoral theses of Lowman (1977) and Bevins (1979), studies which are in general, although not exclusively, biased towards geochemistry and igneous petrology in the context of modern interpretation of palaeovolcanic environments. Lowman's work concentrates on the stratigraphy, geochemistry and petrology of volcanics and sediments to the east of Fishguard; whilst Bevins's work concentrates on the geochemistry, petrology, and metamorphism of the volcanics and associated intrusives to the west of Fishguard (this study area). Both studies are highly meritorious and require consultation.

Based on his doctoral studies Bevins (1978) identified the presence of pumpellyite in the basic igneous rocks of the Fishguard Volcanic Complex, proposing that SW Dyfed had been subjected to metamorphic conditions consistent with the prehnite-pumpellyite facies. Subsequent metamorphic studies (Bevins & Rowbotham 1983, Robinson *et al.* 1980, Robinson & Bevins 1986, Bergström 1980, Savage & Basset 1985) indirectly concerned with metamorphism in the Abereiddi Bay – Fishguard – Prescelly Hills region, are in general agreement that conditions have not exceeded the greenschist facies.

The thick and varied sedimentary sequences which occupy much of the coastal districts have received little attention. Cox (1915) describes the general stratigraphic features of the geology between Abereiddi Bay and Abercastle, distinguishing Upper Cambrian, Arenig, and Llanvirn sediments, whilst documenting numerous volcanoclastic horizons (i.e. *Murchisoni* Ash, Yns Castell Ash, Llanrian Volcanics). Cox (1930) followed this work by publishing a preliminary note on the stratigraphy between Abercastle and Fishguard, aptly describing the complexity of the geology, by stating; *Difficulty in elucidating the stratigraphy of this complicated tract is due to the presence among these Cambrian strata of rock types which present considerable lithological resemblances to groups in the Ordovician rocks against which the Cambrian strata are faulted in involved fashion, and in the scarcity or complete absence of fossils, and the similarities of lithology, disentanglement of the various groups* *is something of a geological puzzle'*. During this latter study, Cox distinguished a tripartite (acid-basic-acid) subdivision of the Fishguard Volcanic Complex west of Fishguard, and whilst having been amended to follow modern stratigraphic procedure (i.e. Porth Maenmelyn Volcanic Formation, Strumble Head Volcanic Formation Goodwick Volcanic Formation; Bevins & Roach 1979), this division has been used by all subsequent workers (Fig. 1.4.).

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Figure 1.4. Geological sketch map of the Fishguard district *after* Bevins (1982), showing the tripartite subdivision of the Fishguard Volcanic Complex, west of Fishguard; a divison introduced by Cox (1930) and used by all subsequent workers.

1.4. THESIS PLAN

Throughout the thesis emphasis is placed upon the newly erected stratigraphic division (Table 3.1.), this being the framework upon which interpretations and discussions are based. Stratigraphic procedure follows as closely as possible to the lithostratigraphic hierarchy given in Holland *et al.* (1978).

Chapter 2 briefly outlines the general geological features of SW Dyfed. Attention is paid to the present lithostratigraphic and biostratigraphic division of the early Lower Palaeozoic sequence.

Chapter 3 firstly outlines the principle styles of deformation which control the regional geology and goes on to introduces a new stratigraphy for the Fishguard – Porth-gain district.

Chapter 4 is concerned with discussing the sedimentary stratigraphy, documenting lithotype sections, and outlining regional correlation across SW Dyfed. Facies analysis of individual stratigraphic units is presented and likely environments of deposition discussed.

Chapter 5 documents the igneous geology of the area, outlining the volcanostratigraphy of the area and the regional relationships of the various successions. This is accompanied by descriptions of the major lithologies.

Chapter 6 is concerned with metamorphism, quantifying the conditions imposed by using mineral chemistry and secondary mineral assemblage development, in conjunction with experimental phase relations in the model basaltic system. Chapter 7 summarises conclusions drawn from previous chapters and outlines areas for future research.

References made to different parts of the thesis are given in italic brackets, e.g. {3.3.2}. Localities referred to in this text are given eight figure national grid references from the kilometre grid of the Ordnance Survey; all are located on the 1:10000 sheets SM83SE, SM83NE, SM84SE, SM93SW, SM93NW and SM94SW. Point references are given in square brackets e.g. [S.M. 9834 3476]. Unless otherwise stated, the scales on field photographs are show by a lens cap which measures 5cm in diameter, or a geological hammer the shaft of which is 35cm long.

CHAPTER TWO
GEOLOGICAL SETTING

2.1. INTRODUCTION

The geology of SW Dyfed, of which the Fishguard — Porth-gain district forms an integral part, can be broadly divided into two areas. Ordovician sediments, particularly argillaceous successions of Arenig, Llanvirn, and Llandeilo-Caradoc age occupy much of the area north of St. Davids, whilst to the south and east Precambrian volcanics and thick Cambrian clastics are developed (Fig. 2.1.b.). Major Ordovician volcanic activity was centred around the districts of Fishguard, Abereiddi Bay, Trefgarne, and Ramsey Island, whilst contemporaneous intrusives were emplaced at high-levels into the pre-Llandeilo stratigraphy (Fig. 2.1.a.). All successions were subjected to low-grades of metamorphism and deformed during the Caledonian Orogeny.

This following chapter whilst not exhaustive, briefly discusses the salient geological features of the Precambrian to mid-Ordovician (Tremadoc to Llandeilo) sequence of SW Dyfed; particular attention is paid to the biostratigraphic and lithostratigraphic divisions presently in use. Detailed reviews regarding the historical development of Cambrian throughout Wales are given by Stubblefield (1959), Cowie *et al.* (1972) and Rushton (1974); the development of the Ordovician is addressed by Williams (1969). Igneous activity, metamorphism and Caledonian deformation is briefly outlined, detailed discussions of which are given in subsequent chapters.

2.2. SW DYFED GEOLOGY (Precambrian to Mid-Ordovician)

It is has long been inferred that the Lower Palaeozoic successions of SW Dyfed accumulated along the southern margin of the Welsh Basin {1.1.}. This inference which has bearings on aspects other than sedimentation (i.e. metamorphism {6.1.}), is based on several factors, including; the areas present spatial relationship within the Welsh Caledonides (Fig. 1.1.), arcuate trend (NE-SW to E-W) of major Caledonide which may reflect parallelism with the basin margin, the presence of shallow marine sediments, the apparent northerly ('basin-ward') younging of the regional sedimentary stratigraphy, and peripheral Precambrian inliers which fringe the Lower Palaeozoic cover (Fig. 2.1.); all of which can be interpreted to support a 'marginal' shelf setting with regards to a basin situated to the north.

It is convenient for present purposes to assume that such a setting is correct, however, unlike many other areas of the Welsh Basin where shelf-basin and intra-basin morphologies are defined by pre-orogenic lineaments and diachronous shelf-basin facies relationships, a prolonged or discrete shelf margin can not be discerned, whilst the present day distribution of the Lower Palaeozoic sequence almost certainly does not reflect the true palaeogeography.

A

B

Figure 2.1. Geological sketch maps of SW Dyfed, showing the distribution of major stratigraphic units and volcanic centres. Figure (A) *after* Turner (1979), (B) *after* Allen 1982, compiled from various sources.

Precambrian Basement

Remnants of the Precambrian basement crop out in two fragmented E-W trending belts around the St. Davids and Haverfordwest districts (Fig. 2.1.). The successions are generally divided into two principle stratigraphic units, the Pebidian–Dimetian Complex and Johnston Complex, which are further subdivided (see Baker 1982). Both Complexes are of late Precambrian age (U-Pb zircon dates: Johnston Complex 643 Ma. (+5 or −28), Pebidian-Dimetian Complex 587 Ma. (+25 or −14); *c.f.* Patchett & Joycelyn 1979), representing fragments of an extensive cratonic terrain which developed over southern Britain during a prolonged period of island-arc – continental margin activity and metamorphic events between 700 Ma. and 450 Ma.; culminating in uplift and erosion prior to deposition of the Cambrian over much of Wales (Thorpe 1979, Thorpe *et al.* 1984).

The Pebidian–Dimetian Complex which dominates much of the area south and east of St. Davids, is composed of intrusive granophyres and rhyolitic extrusives of varying affinity, including subaerial bedded tuffs, debris flows, and coherent silicic lavas. The Johnston Complex is comprised of heterogeneous intrusives including, gabbros, diorite, tonalites, pegmatitic granites, and various hybrids, although its association within the Lower Palaeozoic sequence of SW Dyfed is uncertain due to involvement in Variscan deformation.

The significance of the Precambrian basement in controlling subsequent Palaeozoic sedimentation is difficult to assess in the light of the structural complexity {3.4.}. It would appear however, that whilst there is little evidence to indicate direct involvement, it is clear that during much of early Lower Palaeozoic times basement rocks were a significant crystalline source, possibly shed from uplifted terrains situated to the south (Crimes 1970); although it is suspected that basin topography varied considerably with local topographic highs developing in response to basin margin tectonics, exposing basement rocks to erosion within the shallow Cambrian and early Ordovician seas.

Cambrian Stratigraphy

Cowie *et al.* (1972) replace the previously held divisions of Lower, Middle and Upper Cambrian, with correspondingly named Series; the Comley, St. Davids and Merioneth respectively. However, the terms Lower, Middle and Upper Cambrian are used in the following text due to the uncertainty regarding the exact age relationship of the various Cambrian successions in the Fishguard – Porth-gain district.

Green (1908) demonstrated the basal Cambrian unconformity in the St. Davids district upon which a complete, although in part poorly understood Cambrian stratigraphy has been erected by subsequent workers (Table 2.1.a.).

***Lower Cambrian** (Caerfai Group):* The onset of Lower Palaeozoic deposition in SW Dyfed is recorded by Lower Cambrian transgressive conglomerates of the Caerfai Group which lie with significant disconformity on crystal-lithic tuffs and coherent rhyolitic lavas of the Pebidian Volcanics. At its type locality south of St. Davids, the Group falls naturally into four members (Table 2.1.a.). A thick polymictic conglomerate passes upwards into coarse lithic arkose sandstones (St. Nons Member), which gives way gradationally to distinctive red marine mudstones (Caerfai Bay Shale Member) and massively bedded coarse red and green sandstones (Caerbwdy Member). Collectively the Group reflect a deepening-up sequence transitional from intertidal to shallow marine shelf sediments (Crimes 1970).

***Middle Cambrian** (Solva & Menevian Groups):* The Middle Cambrian of SW Dyfed is documented by the Solva and Menevian Groups, which represent the development of the stratotype St. Davids Series (Table 2.1.a.). The successions have been discussed by many (e.g. Cox et al. 1930a, Nicholas 1933, Davis & Downie 1964, Williams & Stead 1982) although the majority of research has been of a biostratigraphic nature due to a proliferation of 'post-Caerfai' fauna; trilobites being the main taxa upon which studies are based (see Thomas *et al.* 1984).

Solva Group sediments reflect a thick alternating intertidal/subtidal clastic sequence, which progressively gave way to relatively open wave and current agitated siliciclastic-argillic shelf sediments of the Menevian Group (Fig. 2.2.). A temporary deepening of the environment during the late Middle Cambrian (i.e. Upper Menevian Group, Table 2.1.a.) and early Upper Cambrian (i.e. lower successions in the *Lingula* Group, Table 2.1.a.) is suggested by coarse sandstones and mud-silt sequences interpreted as the products of distal turbidity currents (Crimes 1970, Williams & Stead 1982). However, deposition of sandstones may reflect storm-generated traction currents, whilst the development of thick interbedded laminated mudstones may reflect prolonged periods of 'fair-weather' sedimentation; it being thought here that a muddy shelf facies is more appropriate rather than deep marine conditions.

***Upper Cambrian** (Lingula Group);* Overlying the Menevian Group, locally with conformity (Cox *et al.* 1930a), beds referred to here as the *Lingula* Group (the *Lingula* Flags of previous workers) represent the development of the Upper Cambrian in SW Dyfed. The Group is correlated with the stratotype '*Lingula* Flags' of North Wales, where division into the Maentwrog Stage, Festiniog Stage, and Dolgelly Stage is made. In SW Dyfed no subdivision exists, it being suggested that the majority of beds at outcrop are of Maentwrog age (i.e. lower Upper Cambrian) as beds of Festiniog and Dolgelly age have nowhere been proven (Rushton 1974).

(A)

		Hicks 1881	Stratotype Cambrian in SW Dyfed			
CAMBRIAN	Upper (*Merioneth Series*)	*Lingula* Flags: Tremadoc; Dolgelly Group; Festiniog Group; Maentwrog Group	*Lingula* Group (+ 600m) (Maentwrog Stage)			
CAMBRIAN	Middle (*St. Davids Series*)	Menevian Group	Menevian Group	Upper Menevian beds		c. 30m
				Middle Menevian Beds		c. 100m
				Lower Menevian beds		c. 100m
		LOWER CAMBRIAN: Solva Group	Solva Group	Upper Solva beds		c. 50m
				Middle Solva beds c. 80m	Grey Sandstone Mb.	
					Purple Sandstone Mb.	
				Lower Solva beds		c. 50m
			Unconformity ?			
CAMBRIAN	Lower (*Comley Series ?*)	LOWER CAMBRIAN: Caerfai Group	Caerfai Group	Caerbwdy Sandstone Member		c.150m
				Caerfai Bay Shale Member		c.15m
				St. Nons Sandstone Member		c.140m
				Conglomerate		c.10-15m
			Unconformity			
Precambrian			Pebidian-Dimetian Complex			

			S.W.Dyfed *1	Carmarthen/Llandeilo
ORDOVICIAN	Llandeilo	*Nemagraptus gracilis*	Llandeilo/Caradoc Shales	*N. gracilis* Shales
			Castell Limestone	Mydrim Limestone
	Llandeilo	*Glyptograptus teretiusculus*	Caerhys Shale Formation	*G. teretiusculus* Shales
				Llandeilo Flags
				Ffairfach Group
	Llanvirn	*Didymograptus murchisoni*	Aber Mawr Formation	*D. murchisoni* Shales
	Llanvirn	*Didymograptus artus*		*D. artus* Shales *2
				Llanfellteg Formation
	Arenig	Fennian	Road Uchaf Formation	Pontifenni Formation
	Arenig	Whitlandian	Penmaen Dewi Formation (*Tetragraptus Shales*)	Colomendy Formation
			Abercastle Formation	Afon Ffinnantt Formation
	Arenig	Moridunian		Carmarthen Formation
			Ogof Hên Formation: Shale Mb.; Sandstone Mb.	Ogof Hên Formation: Bolahaul Mb.; Allt Cystanog Mb.
	Tremadoc			Tremadoc Shales

(B)

Table 2.1. (A). Stratotype Cambrian of SW Dyfed (modified after Williams & Stead 1982). (B) Composite Lower Ordovician stratigraphy and principle stratigraphic subdivision in South Wales, complied from various sources (see text): (*1 Composite stratigraphy from Ramsey Island and Abereiddi Bay; *2 *D. artus* Shales = *D. bifidus* Shales of others).

The *Lingula* Group represents the thickest (>600m) development of Cambrian strata in South Wales, characterised by the frequent occurrence of *Lingulella davisii* (M'coy) and intertidal – subtidal thin bedded ('ringers') sandstone/mudstone alternations (Turner 1977). The regional development of the Group and widespread documentation of its 'ringer' lithology, suggests that sedimentation was for the most part above effective storm-wave base; it being visualised that SW Wales was the site of an extensive shallow siliciclastic shelf throughout much of the lower Upper Cambrian {4.3.}.

Ordovician Stratigraphy (Tremadoc to Llandeilo)
In comparison to the Cambrian the development of the Ordovician in SW Dyfed is defined within a far more rigid stratigraphic framework. However, the regional palaeogeography remains for the most part poorly appreciated and biostratigraphic problems remain numerous (Kennedy 1986, Whittington *et al.* 1984)

Tremadoc Series: Hicks (1873,1875) denoted large areas of SW Dyfed as belonging to the Tremadoc Series, although subsequent studies showed Hick's *'Tremadoc Beds'* to contain Arenig faunas (Pringle 1911, 1930). At present, rocks of Tremadoc age having nowhere been proven in SW Dyfed, their absence attributed to uplift and erosion during late Tremadoc times (Fortey & Owens 1978). Evidence for this early Ordovician non-sequence is seen on Ramsey Island where thin conglomerates and shallow marine sandstones of Ogof Hên Formation (i.e. Lower Arenig; Table 2.1.b.) lie disconformably on the Upper Cambrian (Kokelaar *et al.* 1985). To the east around Carmarthen, no major disconformity is observed, in as much as Tremadoc strata are widely developed (e.g. Cope *et al.* 1978). This may suggest that whilst SW Dyfed and SE Dyfed were in close spatial proximity they each had separate sedimentary identities, existing as discrete terrains during early Ordovician times.

Arenig Series: Cox (1916, 1930) stressed in detail the two-fold lithological division of the Arenig in S Wales, comprising a lower arenaceous and upper argillaceous division, the 'Abercastle Beds' and '*Tetragraptus* Shales' respectively (Table 4.4.). Fortey & Owens (1987) whilst accepting this broad lithologic division, have established three stages within the Arenig of South Wales, termed the Moridunian, Whitlandian and Fennian. This subdivision is followed here.

Moridunian Stage: The stratotype Moridunian crops out on Ramsey Island, where lowermost Moridunian sediments of the Ogof Hên Formation rest disconformably on the Upper Cambrian (Bates 1969, Kokelaar *et al.* 1985); the Formation is also identified in the Carmarthen district resting with apparent discordance on the Tremadoc (Fortey & Owens

1978). In both areas the Formation shows similarities, characterised by fauna of the *Merlinia selwynii* Biozone, and transgressive polymictic conglomerates which rapidly pass upwards into mudstones (Fig. 2.2.)

Whitlandian Stage: The type Whitlandian successions are all exposed in the Carmarthen district (see Fortey & Owens 1987). Of interest to this study is the complete revision by Fortey & Owens (1987) of the Arenig stratigraphy in the Abereiddi Bay - Abercastle district, placing both the Abercastle Formation and the Penmaen Dewi Formation (the *Tetragraptus* Shales of Cox 1916) within the Whitlandian (Fig. 2.1.b.); effectively restricting the exposed Arenig of mainland SW Dyfed to the Middle Arenig. A discussion regarding the merits of the present Arenig stratigraphy is given in Chapter 4, although irrespective of stratigraphic interpretation it is clear that during mid-Arenig times the regional depositional system changed from shallow shelf clastics to one in which mudstones accumulated within a relatively deep and possibly restricted marine environment; conditions that were to persist to the late Upper Llanvirn (Fig. 2.2.). The cumulative thickness of the Arenig/Llanvirn argillic pile across SW Dyfed has been estimated to be in excess of 600m (George 1970), although local thickness and regional distribution remains largely uncertain; whilst subdivision is based purely on faunal grounds in the absence of which definition between the Arenig and Llanvirn shales can generally not be made.

Fennian Stage: Shales of Fennian age have been identified on Ramsey Island (Kokelaar *et al.* 1985, Road Uchaf Formation; Table 2.1.b.), although not on the mainland of SW Dyfed. Realistically however, the upper parts of the Penmaen Dewi Formation are likely to be equated, although the faunas required for the unravelling of the 'Arenig shales' of SW Dyfed have yet to be forthcoming. At present, the base of the Llanvirn Series defines the top of the Fennian, however, at Abereiddi Bay the Llanvirn type-locality (see below), no Fennian assemblages have been identified (Fortey & Owens 1987).

Llanvirn Series: The Llanvirn is dominated by monotonous black marine shales, the Arenig - Llanvirn boundary being recognised entirely on faunal grounds, the base of the Llanvirn being marked by an influx of pendant didymograptids. Based on diagnostic fauna, the Llanvirn is classically divided into Lower and Upper divisions, the Lower Llanvirn being characterised by *Didymograptus bifidus* Hall, the Upper Llanvirn by didymograptids attributed to *Didymograptus murchisoni* (Table 2.1.b.). This division is presently proving unsatisfactory due in part to the taxonomic confusion regarding pendant didymograptids, in particular *D. bifidus* (Hall) which is not conspecific to the British Llanvirn (Cooper & Fortey 1982). As a substitute for *D. bifidus* Hall, Fortey & Owens (1987) propose a more distinctive Lower Llanvirn species *Didymograptus artus* as a suitable zonal fauna. This is followed here, although several assemblages documented in this study as belonging to the *murchisoni* Biozone contain *D. artus*, highlighting the need for further revision. Similar difficulties arise out of the continued use of *D. murchisoni* to define the Upper Llanvirn, members of this species having long been recognised in horizons of the '*artus*' Biozone (Kennedy 1986). A substitute has as yet not been put forward and its use is retained.

During the late Llanvirn shallow marine conditions prevailed around Carmarthen and Llandeilo with the accumulation of sandstones, ashes, and shelly mudstones (i.e. Ffairfach Group, Table 2.1.b.). However, thick graptolitic marine mudstones were being deposited elsewhere (i.e. Caerhys Shale Formation, Table 2.1.b.) possibly reflecting a varied basin morphology {4.1.}.

Llandeilo Series: The stratotype Llandeilo Series, developed around the Llandeilo and Llangadog districts of SE Wales, reflects a transgressive sequence of shallow marine sandstones and calcareous mudstones which progressively pass upwards into deeper water sediments (Williams 1953). Division into Lower, Middle and Upper Stages has been suggested (Wilcox & Lockley 1981), although this division is based principally on shelly and trilobite faunas of which a greater understanding is required before it has wider zonal usage (Thomas *et al.* 1984).

In SW Dyfed the majority of Llandeilo successions are graptolitic marine mudstones and laminated muddy carbonates (Fig 2.2.), only around Narbeth are shelly faunas and shallow marine conditions documented (Addison 1974). Llandeilo sediments of graptolitic facies are commonly divided into the *Glyptograptus teretiusculus* and *Nemagraptus gracilis* biozones (Table 2.1.b..). It is customary to relate the *G. teretiusculus* Biozone with the base of the Llandeilo, however, it has been shown that the lower part of the '*teretiusculus*' Biozone is older than the earliest Llandeilo (Bergström 1971, 1973); the problems posed by this are as yet unresolved. The *gracilis* Biozone represents the Upper Llandeilo, although diagnostic faunas also span the Constonian Stage of the overlying Caradoc. Successions containing both biozones, effectively the entire development of the Llandeilo Series, have been identified (Toghill 1970) around Mydrim (i.e. *G. teretiusculus* Shales and Mydrim Limestone: Table 2.1.b..), whilst at Abereiddi Bay it has been suggested (Hughes *et al.* 1982) that both the *teretiusculus* and *gracilis* biozones overlie the type-Llanvirn (i.e. upper parts of the Caerhys Shale and Castell Limestone; Table 2.1.b.).

Igneous activity

During the Lower Ordovician (Arenig - Llanvirn) SW Dyfed was a major centre of igneous activity, with the thick local accumulation of ashes, lavas, and widespread emplacement of

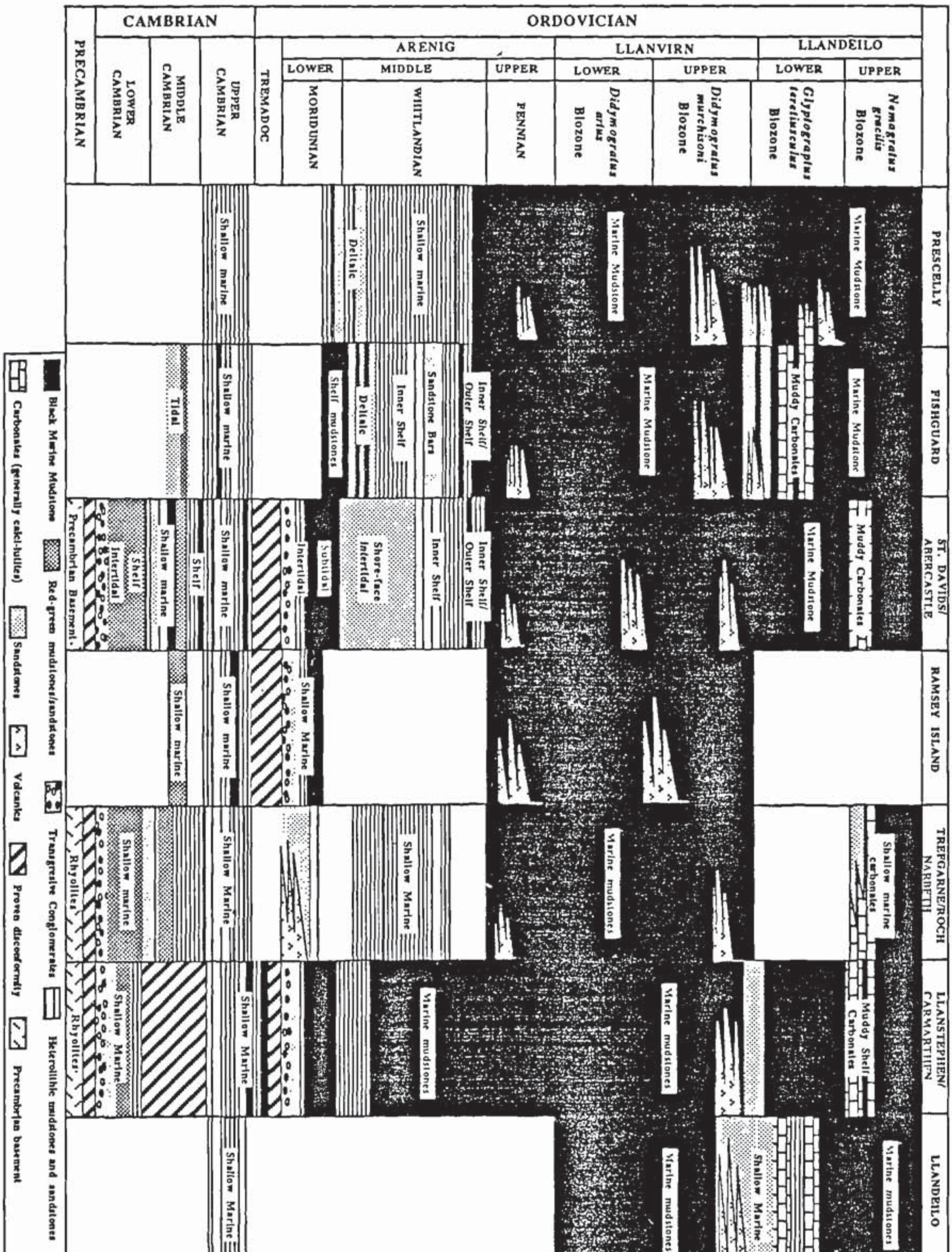


Figure 2.2. Schematic presentation of the early Lower Palaeozoic sedimentary sequence throughout SW Dyfed complied from various sources (see text), including work by the present author which is presented in the following text. The figure also shows the widespread development of volcanics during late Arenig to Upper Llanvirn times in part contemporaneous with the major change in the depositional system from siliciclastic shallow marine shelf sediments to pre-dominantly argillic.

intrusives. The first volcanic episode is documented by the Trefgarne Volcanic Group, a sequence of subaerial andesitic lavas and bedded tuffs, believed to be of Lower Arenig age (Fig. 2.2.). During the upper parts of the Arenig numerous tuffaceous horizons are documented {5.2.}, although an identifiable volcanic source (or sources) is lacking.

During Llanvirn times gabbros, tonalites, diorites, and microgranite sills were emplaced at high levels into the pre-Llandeilo sedimentary cover. Active Llanvirn volcanism is recorded by numerous discrete volcanic centres, the largest of which crop out at Abereiddi Bay (Llanrian Volcanics), Ramsey Island, and throughout the Fishguard – Prescelly district (Fishguard Volcanic Complex). Volcanism occurred for the most part within the marine environment, the depth of which varied and influenced both the style of eruption and nature of resulting deposits. At Abereiddi Bay and on Ramsey Island, volcanism of predominantly silicic composition developed a wide variety of marine volcanic products; including turbidic tuffs, cohesive debris flows, rhyolitic lavas, welded and non-welded ash-flow tuffs, and other mass-gravity flow deposits; all of which appear to have been actively influenced by syn-volcanic tectonism (Kokelaar *et al.* 1984a, Kokelaar *et al.* 1985). The Fishguard Volcanic Complex can be viewed as in general character with volcanism elsewhere, although bimodal (acid-basic) magmas resulted in deposits developing in response to compositional control as well as physical processes and environmental constraints {5.1.}.

Metamorphism

Based on the identification of pumpellyite from metabasites within the Fishguard Volcanic Complex, Bevins (1978) recognised that SW Dyfed had been subjected to low grades of metamorphism, suggesting that conditions consistent with the prehnite-pumpellyite facies had be attained. Roach (1969) had previously identified mixed prehnite-stilpnomelane-epidote-amphibole assemblages from the St. Davids - Carn Lliddi intrusion west of St. Davids, the origin of which he attributed to late stage deuteric alteration, although subsequently regarded as of metamorphic origin (Roach & Floyd, *pers comm.*: in Bevins 1978).

Bevins & Rowbotham (1983) subsequently quantified the St. Davids - Carn Lliddi assemblages as belonging to the prehnite-pumpellyite facies, whilst in dolerites from the Prescelly Hills they identify greenschist and pumpellyite-actinolite facies assemblages, attributing this apparent northerly directed prograde sequence (i.e. St. Davids = prehnite-pumpellyite; Prescelly Hills = pumpellyite-actinolite facies) as reflecting the influence of the original shelf-basin morphology.

A predominance of anchizone illite crystallinity with Lower Palaeozoic sediments between Dinas Head and Newgale (Robinson *et al.* 1980), contrasts well with the prehnite-pumpellyite metabasite assemblages. More recently however, in a regional survey of illite crystallinity across the Welsh Basin, Robinson & Bevins (1986) identify only epizone crystallinity in Ordovician sediments north of St. Davids, although appear to give no explanation as to the reason for this apparent discrepancy with the earlier study of Robinson *et al.* (1980).

Conodont thermal maturation index (CAI) of Llandeilo-Caradoc conodont fauna with values of 5 (i.e. 300-350°C, Epstein *at al.* 1977) at Abereiddi Bay (Bergström 1980, Savage & Basset 1985), compare favourably with metabasite and crystallinity data. Whilst all studies appear to possess common ground as regards the upper-T-limit, in as much as there is no evidence to suggest temperatures greater than 300-350°C have prevailed.

Deformation

It has long been recognised that major lineaments and axial traces of large Caledonide folds swing progressively from Central to SW Wales in an arcuate trend from the NE (Caledonide trend) towards E-W (Fig. 2.1.); a trend increasingly viewed as having been inherited from pre-existing 'grain' within the basement (e.g. Anketell 1987). However, the structural relationships and nature of deformation within SW Dyfed remain poorly understood, only brief documentation exists of major structural features such as the definition of major axial traces and larger faults. The paucity of data results from the poor inland exposures and general uncertainty regarding much of the tectonic fabric of the area between Carmarthen and St. Davids. Nevertheless, there appears to be several recurring themes briefly documented within the literature suggesting in part the influence of a common style (or styles) of deformation, which may reflect regional fold-thrust tectonics {see Chapter 3 for details and review}.

CHAPTER THREE
STRUCTURAL STYLES & PROPOSED STRATIGRAPHY

3.1. INTRODUCTION

This chapter briefly documents the major styles of deformation which complicate the stratigraphy and which have bearings on aspects such as metamorphism which is discussed late in this thesis. The chapter goes on to document the principle stratigraphic successions depicted on maps 1-6 and outlines the salient features of individual units in likely chronologic order (Tables 3.1., 3.2.); causative reasons, type localities, and age constraints are given in subsequent chapters.

3.2. PROPOSED STRATIGRAPHY

In the light of detailed mapping the existing stratigraphy of the Fishguard – Porth-gain district is revised. The existing sedimentary stratigraphy (for details see chapter 4) is for the most part abandoned and a new scheme erected. The existing volcanostratigraphy is modified and several small additions made, although the findings of previous workers are generally accepted.

The stratigraphic divisions proposed herein (Table 3.1.) are based principally on local lithostratigraphic correlation, sparse faunal evidence, and correlation with defined Lower Palaeozoic successions from elsewhere in SW Dyfed {2.2}. Regional events in the context of Welsh Basin geology are also taken into consideration. The author acknowledges that the stratigraphy is not definitive given the lack of biostratigraphic control, although it is regarded a conceptually valid lithostratigraphy upon which discussions and interpretation are based.

The proposed chronologic order, likely biostratigraphic range, and nomenclature used in discussing the various stratigraphic units are outlined in Table (3.1.). Table (3.2.) outlines the salient lithological characteristics of individual stratigraphic units and location of type localities. Distribution and relationships of the stratigraphic units are shown on Maps (1-4) where letter abbreviations have been given to sediments at group level (i.e. LG = *Lingula* Group) and letter codes A–H have been given to sedimentary successions at formational level, additional prefixed numbers are given to subdivisions at member level (i.e. FGB = Fishguard Bay Group; G = Lower Town Formation, G_1 = Castle Point Member, G_2 = Pantycelyn Member; H = Dyffryn Formation, H_1 = Drim Member, H_2 = Tref-wrgi Member; I = *N. gracilis* Shales). The various igneous successions are given letter abbreviations (i.e. PDT = Pwll Deri Tuffs). Poor or complete absence of inland exposure frustrates the ability to define more rigidly stratigraphic trends, leading to large areas being collectively mapped at group/formational level, or remain as undifferentiated strata (Maps 1-4). Division at member level is only possible throughout the coastal districts.

System	Series	Stage	Group	Formation	Member
ORDOVICIAN	LLANDEILO	Upper	Fishguard Bay Group (FBG)	*N. gracilis* shales east of Fishguard	
ORDOVICIAN	LLANDEILO	Lower	Fishguard Bay Group (FBG)	Dyffryn Formation (H)	Tref-wrgi Member (H2)
ORDOVICIAN	LLANDEILO	Lower	Fishguard Bay Group (FBG)	Dyffryn Formation (H)	Drim Member (H1)
ORDOVICIAN	LLANVIRN	Upper	Fishguard Bay Group (FBG)	Lower Town Formation (G)	Pantycelyn Member (G2)
ORDOVICIAN	LLANVIRN	Upper	Fishguard Bay Group (FBG)	Lower Town Formation (G)	Castle Point Member (G1)
ORDOVICIAN	LLANVIRN	Upper	*Murchisoni* Shales	Goodwick Volcanic Formation (GVF)	
ORDOVICIAN	LLANVIRN	Upper	Fishguard Volcanic Complex / Fishguard Volcanic Group	Strumble Head (SHFV) Volcanic Formation	Carreg Gybi Member (CGM)
ORDOVICIAN	LLANVIRN	Upper	Fishguard Volcanic Complex / Fishguard Volcanic Group	Porth Maenmelyn Volcanic Formation (PMVF)	
ORDOVICIAN	LLANVIRN	Upper	Arenig – Llanvirn Shales (F)	*D. murchisoni* Shales (F3)	
ORDOVICIAN	LLANVIRN	Lower	Arenig – Llanvirn Shales (F)	*D. artus* Shales (F2) (Bifidus beds)	
ORDOVICIAN	ARENIG	Fennian	Arenig – Llanvirn Shales (F)	Penmaen Dewi Formation (F1) (Tetragraptus Shales)	Pwll Deri Tuffs (PDT); Yns Castell Ash (YCA)
ORDOVICIAN	ARENIG	Whitlandian		Trwyn Llwyd Formation (D)	Ogof Aderyn Member. (D5)
ORDOVICIAN	ARENIG	Whitlandian		Trwyn Llwyd Formation (D)	Ogof Felin Member (D4)
ORDOVICIAN	ARENIG	Whitlandian		Trwyn Llwyd Formation (D)	Godor Mabli Member (D3)
ORDOVICIAN	ARENIG	Whitlandian		Trwyn Llwyd Formation (D)	Aberbach Member (D2)
ORDOVICIAN	ARENIG	Whitlandian		Trwyn Llwyd Formation (D)	Pen Deudraeth Member (D1)
ORDOVICIAN	ARENIG	Whitlandian		Abercastle Formation (E)	Pwll Llong beds (E2)
ORDOVICIAN	ARENIG	Whitlandian		Abercastle Formation (E)	Pen Porth-egr member (E1)
ORDOVICIAN	ARENIG	Moridunian		Porth-gain Harbour Formation (C)	Shale member
ORDOVICIAN	ARENIG	Moridunian		Porth-gain Harbour Formation (C)	Sandstone member
CAMBRIAN	Upper		*Lingula* Group (LG)	Porth Ffynnon Formation (B2)	
CAMBRIAN	Upper		*Lingula* Group (LG)	Mynydd Morfa Formation (B1)	
CAMBRIAN	Middle			Lach Dafad Formation (A)	Pwlldawnau Member (A3)
CAMBRIAN	Middle			Lach Dafad Formation (A)	Tregwent Member (A2)
CAMBRIAN	Middle			Lach Dafad Formation (A)	Carreg Herefio Member (A1)
CAMBRIAN	Lower				

Table 3.1. Proposed stratigraphic subdivision of Lower Palaeozoic sequence in the Fishguard – Porth-gain district (*n.b.* letter codes relate to stratigraphic units depicted on Maps 1-4). The stratigraphic nomenclature of the Fishguard Volcanic Group follows Bevins & Roach (1979a).

In the absence of any evidence to the contrary, contacts between individual formations where not exposed are generally assumed to be conformable; although the likely probability is acknowledged, but cannot be proven, that all inter-formational contacts are of a tectonic nature.

3.3. MAJOR STYLES OF DEFORMATION

The Lower Palaeozoic sequence of the Fishguard – Porth-gain district have been variably affected by complex deformation. In the following section the principle styles of deformation are briefly documented, the appreciation of which has done much to elevate the stratigraphic complexity. The observations presented may be regarded as a synthesis rather than an detailed study; nevertheless the structures which are believed to control the regional distribution of geology may be applicable to other areas of SW Dyfed.

Phases of deformation

It is thought that deformation in the Fishguard – Porth-gain district can be best explained under three broad headings of **a)** pre-orogenic structures, **b)** Caledonian deformation (fold-thrust tectonics), and **c)** Caledonide cross-faults ((NE-SW)). A further set of cross-faults with a NW trend are evident locally and may reflect the effects of Variscan deformation, given the close proximity of the Variscan Deformation Front to the south of this area (Fig. 1.2.), the significance of such structures is only locally apparent, although it is thought that they may have a substantially greater control on the geology than is depicted on Maps 1-6.

a) Pre-orogenic structures

'Soft sediment deformation' (*sensus* Maltman 1984), and deformation integrally associated with the igneous geology of the district (i.e. 'syn-igneous') may locally complicate the stratigraphy, although are not thought to have regional implication. Only small scale features have been identified during this study, although the importance attached to this style of deformation in both volcanic and sedimentary terrains within the Welsh Basin is increasingly recognised (e.g. Kokelaar *et al.* 1985, Fitches *et al.* 1986). It seems likely that large pre–Caledonide structures would have existed in controlling sedimentation and volcanism within this and adjoining areas of SW Dyfed, although at present direct evidence is lacking.

'Soft-sediment' Deformation: Small scale 'soft-sediment deformation', whilst not prevalent, is not uncommon to the thick sedimentary sequence. Apart from structures such as convolution and microfolding of lamination (see Plate 4.3.), there is evidence to suggest that sediments where locally disrupted by gravity induced slumping and non-sedimentation processes.

Slumping of sedimentary units whilst not common, is locally seen to be of considerable magnitude; excellent examples of which are seen within the newly introduced Trwyn Llwyd Formation, north of Godor Mabli [S.M. 8810 3650], where the succession at this point is interpreted to reflect the transition from deltaic sandstones (Plate 3.3.) into thick siliciclastic-argillic shelf sediments {4.4.}. The slumps in question occur as sheets which show a high degree of lateral continuity, being variable in thickness from 10's of centimetres to metre-scale units, internally exhibiting bedding planar overfolds and extensive convolution. Locally, overfolds are seen as detached 'fish-hook' folds, isolated in homogenised coarse lithic-arenites (Plate 3.1.); such deposits may reflect relatively small displacements, although it is conceivable that entire sheets may have moved considerable distances en masse (Pickering 1982).

'Syn-igneous' deformation: Evidence for small scale pre-orogenic deformation associated with the igneous geology of the area can be seen throughout the Fishguard Volcanic Group. Small syn-sedimentary faults are common to thin intervening sedimentary partings, the origin of which may reflect seismic activity during volcanism and periods of quiescence. Fluidization (*c.f.* Kokelaar 1982) of unconsolidated, or poorly lithified sediments. by high-level intrusives, whilst not prevalent is not uncommon; although the forces here are not from the sedimentation process. Examples of the interaction of magma and wet- or poorly-lithified sediments are given in Chapter 5. The extent to which intrusion and accompanying deformation affected the host sediment is difficult in many cases to assess as magma/sediment interactions are generally confined to argillic sediments which have little or no stratification. However, from observations of specific hypabyssal sills, it has been possible to trace individual mud/magma associations over strike length greater than 4km {5.4.}, suggesting whole sedimentary horizons were thixotropically disturbed.

The recent suggestion by Kokelaar *et al.* (1984b) of large scale penecontemporanueous wet-sediment slides within the Goodwick Volcanic Formation and surrounding strata appears in part, to have been invoked as a way of accounting for complex stratigraphic relationships. The present author believes that such relationships are thought better interpreted as Caledonide thrusts in character with the structure elsewhere; being essentially contractional rather than extensional structures, although this is not to infer inversion.

b) Principle Phase of Caledonide Deformation

Stratigraphic complexity throughout the Fishguard – Porth-gain region is thought to relate to one principle phase of Caledonide deformation which has a fold-thrust signature; during which, folding and thrusting are viewed as part of the same event, although this is not to suggest that each phase is contemporaneous throughout the area.

The regional trend of structures associated with this phase of deformation appears to be

E–W, in character with successions to the west at Abereiddi Bay (Waltham 1971) and east towards Newport (Lowman & Bloxam 1981); although subsequent cross-faults complicate both the stratigraphy and pre-existing structure by block rotation (see below). The principle control on the development of structure throughout the area appears to be lithology and competence contrast. Alternating sandstones and mudstone often exhibit spectacular local buckling, whilst argillic lithologies show extensive cleavage planar movement, although little in the way of folding. Igneous rock types act as essentially rigid blocks accommodating strain either by low amplitude undulatory buckles or thrusting along ductility contrasts such as interbedded mudstones and hyaloclastites.

It is assumed that thrusts, folds and cleavage are late Silurian/early Devonian in age; there is no evidence to indicate that structures (other than those of 'soft-sediment' origin) within the Cambrian and Ordovician strata of the Fishguard – Porth-gain district are older than structures seen in younger Caledonian strata elsewhere. However, the presence of apparently unrelated large Caledonide cross-faults, complicates age relations in as much as two discrete episodes of Caledonide tectonism appear to be present. The affects on the regional trend by later cross-faulting is seen by 'drag' and block rotation, typified in the area around Pwlchrochan [S.M.8840 3645] , where well defined folds are rotated through 90° with axial traces trending perpendicular to each other in the same section. (Map 3.). Similar rotations are seen through the coastal district and clearly any inland geology must at regarded at best as inferred, whilst the significance of small scale structures must await further study; it requiring a detailed investigation in an attempt to 'strip-away' these apparently later structures, so as to unravel the incongruent relationships.

Folds

Throughout the area folding varies considerably in both nature and scale, from gentle to isoclinal, from *regional folds* to local disharmonic *buckling* . Cleavage, accompanying folding, is developed as predominantly steep to upright varieties of slaty cleavage in argillic lithologies and as an irregular fracture cleavage with arenaceous and volcanoclastic lithologies, reflecting a grain-size dependency which can be commonly observed as refraction patterns in interbedded sandstone-mudstone alternation. On the local scale cleavage is axial planar, the present author has seen no evidence for transecting cleavage. A flat lying, or gently southerly dipping, widely spaced crenulation cleavage is locally evident [S.M. 8800 3660] the significance of which is uncertain, although similar cleavage appears to have been documented at several localities throughout the Lower Palaeozoic successions of Wales (e.g. Fitches 1972).

Plate 3.1. *'Soft-sediment deformation'*. Detached soft-sediment fold, isolated with homogenised coarse sandstone from slump sheets within the Godor Mabli Member, directly overlying deltaic sediments of the Aberbach Member [S.M. 8822 3650] (see Plate 3.6.). The width of the match box used for scale (arrowed) is 5cm.

Plate 3.2. Disharmonic folds in thin bedded sandstone-mudstone alternations of the Trwyn Llwyd Formation (Godor Mabli Member) on the headland of Carreg Golfcha [S.M. 8840 3520). Note 'suspended' anticline limb in top right hand corner (arrowed). Such structures are common and reflect small scale thrust accommodation; the sense of movement from left to right, in the field from north to south. The hight of the cliff is approximately 15m.

Regional folds

The axial traces of the principle regional folds, the Llwanda Anticline and Goodwick Syncline (*c.f.* Cox 1916, Thomas & Thomas 1956), grossly dominate the area from Fishguard to the Pen Caer Peninsular (Fig. 1.3.), although extensive disruption by cross-faulting is widely evident (Map 1–4.). It is only the distribution of the Fishguard Volcanic Group at outcrop that allows such structures to be broadly discerned, whilst variable dips commonly opposing the general sense of an anticline-syncline pair are common, and attributable in part to subsequent cross-faults. The paucity of data on both structures is exemplified by the fact that throughout the entire southerly limb of the Llwanda anticline no sediments are seen to be exposed, whilst the presence of the Goodwick Syncline can only be inferred due to the Goodwick Fault, apparently occupying the position of synclinal axis (Map 1.). Departure from a classic anticline-syncline pair is widely evident, exemplified in the Carn Gelli (Maps 1&2.) districts where outcrop is sufficient to delimit small open folds developed in relatively flat bedded silicic volcanoclastics; clearly not coincident with steeply dipping fold limbs. Nevertheless, both the Llwanda Anticline and the Goodwick syncline appear to be valid structural units. It is uncertain as to the relationship of the anticline-syncline pair with both local buckling and regional folds elsewhere, they may relate with the larger Velendre Anticline to the east (Evans 1945), or reflect fold-thrust structures developing in response to the anisotropy between the Fishguard Volcanic Group and the relatively ductile underlying sediments.

Local Folding (Buckling):

Folding at outcrop scale is extremely variable, the style of which is strongly dependant on anisotropy contrasts, particularly bedding thickness and competency (Plate 3.4.) Thin bedded mudstone-sandstone alternation (e.g. *Lingula* Group – Upper Cambrian; Godor Mabli Member – Middle Arenig) locally display spectacular disharmonic and chevron folds, whilst competent lithologies (i.e. Yns Castell Ash, Map 4.) generally display long wavelength low amplitude buckles; stratigraphic depth appearing to have little or no significance. Where block rotation is minimal, axial traces trend E–W and are generally upright (Plate 3.4.), although apparently variably plunging; other than in disharmonic folds there is little evidence for overturning as is common further to the east (Waltham 1971). Fold profiles are invariably asymmetric, northfacing limbs are generally upright whilst southfacing limbs dip at lower angles.

However, folding is not evenly distributed, and not singularly related to anisotropy contrasts. The most intense areas of folding occur in direct association with thrusts, in as much as where folding 'locked-up', further shortening was accommodated by thrusting (Plate 3.3). It is thought that such a process readily accounts for complex field relationships (see Plate 3.5. & 3.6.) and the stratigraphic repetition (Maps 3,4,5,6) where marker horizons are sufficiently preserved (Maps 2-4).

Thrusts

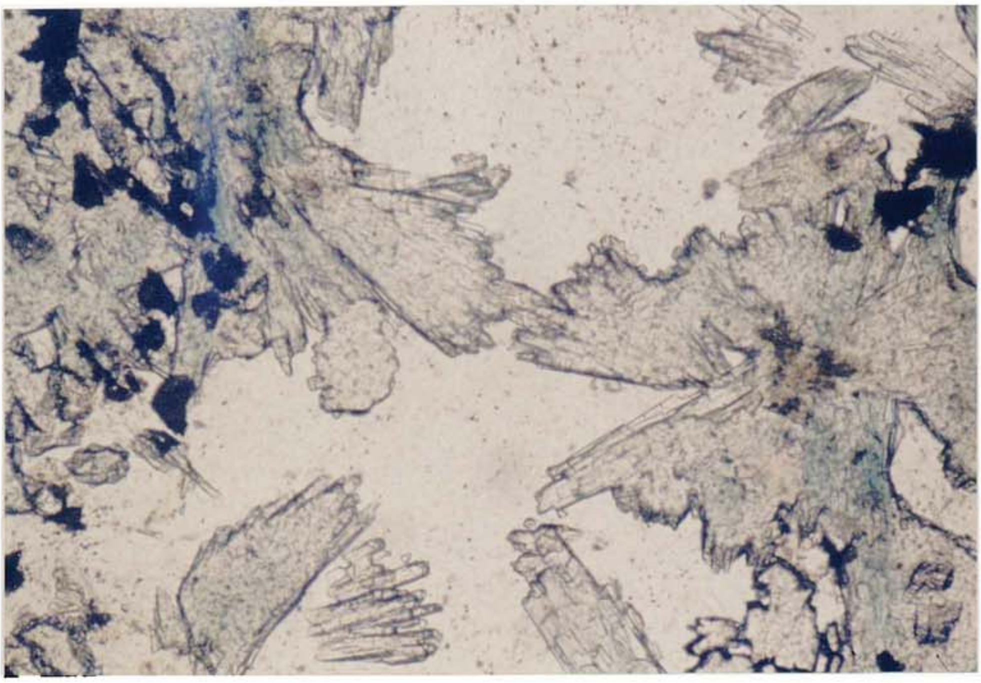
Thrusts occur throughout the area on all scales and throughout the stratigraphy (Plate 3.3., 3.4.), it being the recognition of the existence of such structures which cause the apparent complexity within the stratigraphy. Thrust structures may, or may not, obey the rules of thrust geometry (Dahlstrom 1970, Butler 1982) and cut-up stratigraphic section, although a 'layer-cake' stratigraphy is not thought appropriate, whilst the majority of thrusts are viewed as accommodation structures directly associated with folding. The direction of thrust movement appears to be always from north to south.

The style of thrusting is variable. Some are undoubtedly bedding planar and 'stair-case' dislocations in their own right preferentially confined to inherent weakness such as interbedded pelitic sediments in the Fishguard Volcanic Group. Natural fold-thrust structures are common on the outcrop scale, developing as a direct result of accommodation in fold cores, whilst the subaerial plunging pahoehoe of Thomas & Thomas (1956) is thought better interpreted as a small tip-line fold (Williams & Chapman 1983). However, it is not thought that any of the principle thrusts which control the geology, represent regional décollements, although they may reflect branching off a main surface at depth; possibly the Precambrian basement given its presence at surfacial levels to the east, and apparently shallow basement-cover relationships throughout SW and SE Dyfed (Cope 1979).

Fault-rock products: Common to many thrust surfaces is the preservation of a thin fault-rock which highlights the major plain of dislocation. Such products are best preserved in the Fishguard Volcanic Group, particularly in the vicinity Porth Sychan and Carreg Gybi, where fault rocks vary from the brittle to the ductile field in part dependant on the host lithology. Cataclastites are developed in pillow lavas and dolerites, whilst low-grade mylonites are seen within interbedded sediments, although of specific interest here is not the mechanism, rather the wide variety of low grade secondary minerals developed in such rock types. Basaltic cataclastites commonly contain, prehnite, sphene, epidote, stilpnomelane, pumpellyite (Plate 3.3., 6.4.), whilst sedimentary hosted fault rocks contain clinozoisite, prehnite, neocrystic albite, and traces of white mica; in part reflecting a host rock compositional control, although more importantly their presence indicates that metamorphism and Caledonide deformation were in part contemporaneous. However, it is not thought that phase development and conditions of metamorphism are the direct consequence of strain-heating associated with cleavage development (Roberts & Merriman 1985), although equally it is thought that a burial metamorphic origin can be singularly appropriate (Bevins & Rowbotham 1983, Robinson & Bevins 1986). The observation of cataclastites containing low-grade mineral assemblages,

Plate 3.3. Large scale listric thrust developed at Porth Leuog [S.M. 9050 4110]. The thrust surface runs through the cliff from left to right (arrowed) above marine shales, causing stratigraphic repetition of the Carreg Gybi at outcrop (see Map 6 cross section F to F'). Most of the cliff section is occupied by pillow lavas of the Strumble Head Volcanic Formation (P), overlain by a silicic pyroclastic debris flow (C) which forms the base of the Carreg Gybi Member {5.4.}.

Plate 3.4. (FG22). Photomicrograph of pumpellyite from a coarse cataclastite at Pen Capel Degan [S.M. 9100 4105]. The presence of low-grade minerals is common to all fault-rock products within the Fishguard Volcanic Group, indicating that metamorphism and deformation w ere in part contemporaneous Field of view = 1.7mm



indicate that metamorphism across Wales can not be umbrelled under one model, in as much as metamorphism in the Fishguard district reflects an interaction of processes and events which were pre-, syn- and post-Caledonide deformation. Whether or not the observed grades are the result of, burial, extensional tectonics and high heat-flow, orogenic tectonics, or a combination of all, is another debate which can not be simply resolved {6.1.}.

C). Cross-Faults (Caledonide (NE–SW))

Whilst the principle phase of Caledonide deformation is viewed here in terms of fold-thrust tectonism, it is apparent that such relationships are further complicated by vertical and sub-vertical faults (Maps 1-6), the age of which may be both Caledonide and Variscan, in as much as an array of cross-faults are evident, and no one trend appears totally applicable.

On wave polished surfaces, small cross-faults are generally seen as metre scale shears zones, showing anastomosing pattern with intervening remnant lenses of relatively undeformed rock. The intrinsic nature of the majority of such structures can in part be shown by the lateral continuity offered by a high-level dolerite sill within the Strumble Head Volcanic Formation, which can be traced intermittently through from Pen Capel-Degan [S.M. 9100 4100] eastwards to Maen Jaspis [S.M. 9400 4075], a collective strike length greater than 3km. However, major cross-faults such as the Goodwick Fault are typically Caledonide trending NE-SW structures, the age of which can in part be substantiated by petrographic observations of cataclastites in similar orientated faults from Pwll Arian [S.M. 8850 4030] which contain low-grade Ca-Al silicate mineral assemblages similar to those observed from the fault rock products associated with the early phase of thrusting.

The sense of movement on the majority of structures is not readily discriminated, requiring further study. Locally structures have a predominant dextral sense of movement, although the major faults appear to have a significant dip-slip or oblique component. However, the present author believes that the importance of such structures is comparatively limited, in as much they cross-cut a pre-existing E–W trending Caledonide fabric which is the principle control on the regional geology; these later structures only complicate what appears to have been an already complex picture.

3.4. PRINCIPLE STRATIGRAPHIC DIVISIONS

The sedimentary sequence throughout the Fishguard – Porth-gain district is argued to range in age from the Middle Cambrian to the Upper Llandeilo. Precambrian basement rocks are not observed, although their detrital influence as crystalline source rocks can be seen in many of the clastic successions. The development of the Lower Cambrian (i.e. Caerfai Group {2.3.}) appears to be absent.

The volcanics and associated intrusives which dominate the geology of the area, are thought to reflect two igneous episodes. The first episode is believed to be restricted to the late Arenig, with the intermittent development of silicic pyroclastics, volcanoclastics, and localised high-level emplacement of a suite of felsic intrusives. The major phase of igneous activity is recorded during the Upper Llanvirn, with the development of the Fishguard Volcanic Complex and emplacement into the pre-Llandeilo stratigraphy of generically related basic and intermediate sills.

3.4.1. Principle Sedimentary Divisions

The sedimentary stratigraphy have been totally revised and much of the division is newly introduced (Table 3.2.). It has been found possible to differentiate the successions into 7 principle lithostratigraphic units, referred to here as the Lach Dafad Formation, *Lingula* Group, Porth-gain Harbour Formation, Abercastle Formation, Trwyn Llwyd Formation, Arenig–Llanvirn Shales and Fishguard Bay Group, all of which are further subdivided (Table 3.1.). Informal subdivision is also presented in an attempt to aid communication (i.e. Pwll Llong beds).

1. Lach Dafad Formation (A) – *Middle Cambrian (?)*

The term Lach Dafad Formation is introduced here to collectively describe tectonically fragmented shallow marine red-green sandstone-shale successions and associated strata which crop out at numerous localities between Pwllstrodur and Pwlldawnau (Maps 3 & 4). The Formation is believed to be of Middle Cambrian age {4.2.}, possibly the stratigraphic equivalent of the Solva Group developed south of St. Davids. The general character of the Formation is pieced together from three fault 'isolated' sequences referred to as the Carreg Herefio Member, Tregwent Member, and Pwlldawnau Member, the relationships and stratigraphic attitudes of which are made by conjecture (Table 3.1.). The relationship of the Lach Dafad Formation to younger strata is unknown.

Carreg Herefio Member *(A1):* The Carreg Herefio Member represents a succession of shallow marine coarse grey sandstones and subordinate mudstones which pass conformably into variegated red-green sandstones and marine shales (Table 3.2.), occupying the coastal district between the headlands of Carreg Dandy [S.M. 8815 3553] and Carreg Herefio [S.M. 8820 3607]. The lower grey sandstones are comprised of large scale tabular and trough cross-bedded units and occasional flaser and wavy bedded mudstone partings, the latter reflecting prolonged interludes of 'slack-water' sedimentation in what is otherwise a high energy tidal sequence {4.2.}. The red-green coarse sandstones and red marine shales show a gradational fining-up from thick planar bedded units with shale partings, into finely laminated

Table 3.2. Lithostratigraphic table of defined successions in the Fishguard – Porth-gain district, outlining salient lithological characteristics, location of type localities, and likely environments of deposition.

Systems, Series, Stages, Biozones, and Stratotype Groups				MAJOR STRATIGRAPHIC DIVISIONS (Groups & Formations)	SUBDIVISIONS	Approximate Thicknesses	Location of Type Locality	LITHOLOGY
ORDOVICIAN	LLANDEILO	Upper	*gracilis* Biozone	FISHGUARD BAY GROUP: *N. gracilis* Shales (east of Fishguard)	No Subdivision	> 1200m		marine graptolitic mudstone (see; Lowman 1977, Lowman & Bloxam 1981)
		Lower	*teretiusculus* Biozone	FISHGUARD BAY GROUP: DYFFRYN FORMATION	Tref-wrig Member	> 550m: > 200m	S.M. 9830 5468	**Dyffryn Formation** Graptolitic marine shales: Drim Member composed of calcerous mudstones, passing into black marine shales of the Tref wygl Member. Relatively deep stagnated environment; abyssal depths are not envisaged
					Drim Member	> 350m	S.M. 9329 3729	
	LLANVIRN	Upper	*murchisoni* Biozone	FISHGUARD BAY GROUP: LOWER TOWN FORMATION	Pantycelyn Member	> 220m: > 80m	S.M. 9535 3772	**Lower Town Formation.** Black marine shales (background), interbedded with coarse sandstones, polymictic extraformational paraconglomerates, and epiclastics deposited by sedimentary gravity flows.
					Castle Point Member	>120m	S.M. 96103780	
				FISHGUARD VOLCANIC COMPLEX: *D. murchisoni* Shales; PRESCELLY VOLCANIC GROUP; FISHGUARD VOLCANIC GROUP	Fishguard Volcanic Group: Goodwick Volcanic Formation	> 1500m: > 120m	S.M. 9450 4050	Subaqueous silicic and basaltic volcanics. Pillow lavas, pillow breccias minor hyaloclastites, gabbros, dolerites. Obsidian Flows, rhyolite breccias, density flows, tonalites and microtonalites. Inter bedded pelitic and volcanoclastics
					Strumble Head Volcanic Formation: Carreg Gybl Member	c 40m	S.M. 9020 4120	
					Strumble Head Volcanic Formation	20 – 1200m	S.M. 8970 4130	
					Porth Maenmelyn Volcanic Formation	40 130m	S.M. 8980 3925	
				ARENIG — LLANVIRN SHALES: *D. MURCHISONI* SHALES	No Subdivision	> 100m	No type section	**Arenig-Llanvirn Shales.** Graptolitic black marine shales, subdivision only possible with faunal evidence. Deposition by mud silt turbidites and suspension processes within a deep and possibly restricted marine environment. Local isolated 'distal' ash turbidites with no apparent scource occur at numerous horizons.
		Lower	*artus* Biozone	ARENIG — LLANVIRN SHALES: *D. ARTUS* SHALES (Bifidus Beds)	No Subdivision	? > 100m	No section	**Pwll Deri Tuffs (>40m; S.M. 8900 3820)** Subaqueously emplaced crystal lithic tuffs and 'welded' pyroclastic ash flows, interbedded with marine shales of the Penmaen Dewi Formation. Emplacement below effective storm wave base, althought above the PCL (= 500m)
	ARENIG	Fennian		ARENIG — LLANVIRN SHALES: PENMAEN DEWI FORMATION, *Tetragraptus* Shales; Pwll Deri Tuffs & Ynys-Castell Ash	No Subdivision	> 130m	No type section.	**Ynys Castell Ash (>30m; S.M. 8515 3390)** Thick bedded crystal lithic volcanoclastic mass flow deposits. Contemporenous Arenig intrusive activity locally evident.
		Whitlandian		TRWYN LLWYD FORMATION	Trwyn Llwyd Fm.: Ogof Aderyn Member	> 400m: > 60m	See below	**Trwyn Llwyd Formation.** Prodelta muds passing into deltaic sandstone bodies and shallow marine inner shelf sediments of the *Cruziana* association with periodic emergent sandstone bodies
					Trwyn Llwyd Fm.: Ogof Felin Member	c. 45m		
					Trwyn Llwyd Fm.: Godor Mabli Member	> 160m		
					Trwyn Llwyd Fm.: Aberbach Member.	30-60m		
					Trwyn Llwyd Fm.: Pen Deudraeth Member	> 70m		
				ABERCASTLE FORMATION	Abercastle Formation: Pwll Llong beds	> 70m: c.20m		**Abercastle Formation** Transgressive sedimentation; Shoreface/subtidal sandstones passing into heterolithic inner shelf sediments and sublittoral shelf mudstones
					Abercastle Formation: Pen Porth-egr member	>50m		
		Moridunian		PORTH-GAIN HARBOUR FROMATION	PORTH-GAIN HARBOUR FROMATION: Shale member	> 40m: > 25m	S.M. 81403265	**Porth-gain Harbour Formation** Polymictic breccias and conglomerates discordant upon the Upper Cambrian, passing into heterolithic tidal-subtidal wavey, lenticular and flaser bedded mudstone-sandstone alternations (Sandstone member). Heterolithic's pass gradationally into subtidal mudstones (Mudstone member). Transgressive sequence reflecting deposition probably during the basal Arerig transgression.
					PORTH-GAIN HARBOUR FROMATION: Sandstone member	15m		
							Disconformity	
CAMBRIAN	Upper	Dolgely		LINGULA GROUP (Disconformity)	*LINGULA* Group: Porth Ffynnon Fomation	> 250m: >130m	S.M. 81403265	***Lingula* Group** Thin bedded heterolithic mudstone-siltstone-sandstone alternations. Storm and wave influenced muddy shelf and subtidal environments. Deposition during 'storm' and 'fair' weather
		Festiniog						
		Maentwrog			*LINGULA* Group: Mynydd Morfa Formation	> 250m	S.M. 8730 3480	
	Middle	Menevian Gp.			Lach Dafad Formation: Pwlldawnau Member	> 120m: > 70m	S.M. 8956 4324	Red-green varigated shallow marine inter-tidal sandstones and shales, minor volcanoclastics/epiclastic debris flow
		Solva Group		LACH DAFAD FORMATION	Lach Dafad Formation: Tregwent Member	> 55m	S.M. 8945 3496	Red shallow marine shales passing into laminated heterolithics
					Lach Dafad Formation: Carreg Hereflo Member	>35m	S.M. 8825 3896	Coarse tabular and trough cross bedded tidal sandstones passing into planar bedded red-green sandstones and shales
	Lower	Caerfai Group						

Black marine shales; Igneous lithologies; Marine 'red-beds'; Coarse sandstones; Calcerous mudstones; Heterolithic sandstones/mudstones

Type Locality for Abercastle Formation (Pen Porth-egr member S.M. 8030 3255; Pwll Llong beds S.M. 8415 3330); Type Localites for Trwyn Llwyd Formation (Pen Deudraeth Mb. S.M. 8825 3490; Aberbach Mb S.M. 8825 3490; Godor Mabli Mb. S.M. 8826 3656; Ogof Felin Mb. S.M. 8785 3665; Ogof Aderyn Mb. S.M 8795 3670.)

buff-red siltstones, mudstones.

Tregwent Member *(A2):* The term Tregwent Member is introduced for a distinct, although poorly exposed succession of unfossiliferous red marine shales which pass upwards into thin laminated heterolithic red-green sandstone-mudstone alternations, cropping out in the vicinity of Tregwent Wollen Mill [S.M. 8945 3496] and Treseisyllt Hill [S.M. 8915 3517]. The red shales are characterised by their fine grain size and vivid red colouration, although are generally nondescript, sedimentary structures being limited to fine planar lamination. The overlying laminated siltstones and fine- to medium sandstones, exhibit limited sedimentary structures which suggest deposition within a relatively shallow marine environment {4.2.}, although poor exposure precludes any detailed interpretation.

The red shales of the Tregwent Member have been documented by Cox (1930), who suggested that they may represent the development of the Lower Cambrian Caerfai Group, although the present author suggests that they are likely to represent part of a fragmented Middle Cambrian 'red-bed sequence'; the Lower Cambrian is believed to be absent.

Pwlldawnau Member *(A3):* The term Pwlldawnau Member is introduced here to describe a conspicuous succession of interbedded red-green mudstones, sandstones, granulestones, and volcanoclastics/epiclastics that crop out within the vicinity of Pwlldawnau [S.M. 8800 3700]. The Member shows a complex interplay of sedimentary facies all of which appear to be of shallow marine affinity {4.2.}. The lowermost successions are dominated by bioturbated, buff-red, weakly laminated mudstones/siltstones and occasional fine sandstones, which pass gradationally into thin bedded green-red mudstones, sandstones and subordinate granulestones. A thick (≈ 9m) non-structured volcanogenic debris flow occurs towards the top of the Member, the origin of which is uncertain, although it is believed to represent direct evidence for Cambrian igneous activity in this part of Wales.

2. *Lingula* Group (LG) – *(Upper Cambrian)*

The term *Lingula* Group is introduced for successions which have previously been referred to as '*Lingula* Flags'; the use of the term *'Lingula'* being retained as a well recognised collective term for sediments of likely Upper Cambrian age. Division into the Porth Ffynnon Formation and Mynydd Morfa Formation (Table 3.1.) is based on lithologic variation and the stratigraphic position of the Porth Ffynnon Formation at the top of the Cambrian as developed in this part of SW Dyfed (see below). The Group is thought to represent the development of the Maentwrog and possibly Festiniog Stage of the Upper Cambrian; the development of the Dolgelly Stage is believed to be absent {4.3.}. Collectively, the Group is dominated by shallow marine siliciclastic-argillic shelf sediments characterised by amalgamated sequences of alternating fine-medium bedded mudstones, siltstones, and fine sandstones, within which individual beds are commonly continuous throughout outcrop giving the lithology a conspicuous ribboned ('ringer') appearance; a feature noted by many early workers (e.g. Hicks 1881, Thomas & Jones 1912, Williams 1934).

***Mynydd Morfa Formation** (B1):* The term Mynydd Morfa Formation is introduced for heterolithic shallow marine clastics of the *Lingula* Group which occupy the northern most part of the Mynydd Morfa headland [S.M. 8750 3455] and crop out locally elsewhere (Map 3.). The Formation's typical lithology is one of thin to medium bedded mudstone-siltstone-sandstone and rare granulestone alternations within which a complete array of lithologies and bed forms are developed; varying from coarse sublittoral sheet sandstones through various delicate stacked permutations of flaser, wavy, and lenticular bedded streaked mudrocks, collectively reflecting deposition within a rapidly fluctuating tidal – subtidal environment {4.3.}. The Mynydd Morfa Formation differ from the Porth Ffynnon Formation to the west in as much as it contains a considerably greater percentage of discrete mudstones, whilst it is also suggested, to reflect deposition within a slightly shallower marine environment and occupy a lower stratigraphic position within the Upper Cambrian {4.3.}

***Porth Ffynnon Formation** (B2):* The term Porth Ffynnon Formation is introduced to describe sediments of the *Lingula* Group, which occupy the coastal districts between Porth-gain and Trwyn Llwyd [S.M. 8325 3290] cropping out in near continuous 4km E-W trending belt (Map 4.). The Formation is overlain with hiatus by Ordovician sediments, strongly suggesting (allowing for indeterminate erosion), that its occupies a position near the top of the Upper Cambrian of SW Dyfed. The Formation's typical lithology is one of laterally persistent (>200m) thin to medium bedded, fine- to medium sandstones separated by weakly laminated siltstones and rare mudstones. The sandstones exhibit an array of internal wave and current generated structures, recording discrete high energy events during prolonged periods of silt and mudstone deposition; collectively reflecting an environment visualised as a muddy marine shelf dominated by 'fair-weather' sedimentation periodically agitated by 'storm' related events {4.3.}. Such salient 'ringer-like' features are however, characteristic of siliciclastic Arenig strata of the thick (> 375m) Trwyn Llwyd Formation {4.5.}, making inland correlation in the light of the structural complexity almost impossible without faunal evidence. This points to an area of complexity hitherto unappreciated in SW Dyfed, in as much as, many of the thick successions documented as '*Lingula* Flags' north of St. Davids (Williams 1934, Rushton 1974), may well be of Arenig age.

3. Porth-gain Harbour Formation (C) – *Lower Arenig (Moridunian Stage)*

The term Porth-gain Harbour Formation is introduced for sediments thought to be of lowermost Arenig age (i.e. Moridunian Stage), which lie with hiatus upon the Porth Ffynnon

Formation (i.e. Upper Cambrian); the intervening Tremadoc (and possibly late Upper Cambrian) is believed to be absent. At outcrop the Formation is only observed within a 40m interval at east side of the entrance to Porth-gain Harbour [S.M. 8140 3265, Map 4.], although the interpretation given to this locality is important with regards the understanding of the Arenig stratigraphy and early Ordovician palaeogeography within this study area. The Formation concomitantly fines upwards from intertidal channeled sandstones and polymictic breccias-conglomerates into wavy, lenticular and flaser bedded mudstone-sandstone alternations (***Sandstone member*** Table 3.1.). This lower clastic sequence gives way to subtidal mudstones (***Mudstone member*** Table 3.1.); interpreted here as reflecting steadily deepening marine conditions developing in response to the basal Arenig transgression {4.4.}.

4. Abercastle Formation (E) – *(Whitlandian Stage)*

The Abercastle Formation represents the development of Whitlandian Stage sediments (Fortey & Owens 1987) between Abereiddi Bay and Abercastle. The Formation as defined by previous workers is modified, details of which are given in Chapter 4. The Formation is tectonically fragmented, although its sedimentary character reflects a steadily 'deepening' clastic sequence. Informal subdivision into the ***Pen Porth-egr member*** and ***Pwll Llong beds*** is presented in an attempt to aid discussion regarding internal relationships, whilst it also distinguishes a lower clastic and upper argillic division (Table 3.2.).

The Pen Porth-egr member is dominated by planar, tabular, and trough cross-bedded fine sandstones of likely shoreface origin, passing upwards into fine-medium planar and wavy bedded coarse sandstone-mudstone alternations exhibiting limited features characteristic of a shallow siliciclastic shelf {4.4.}. The Pwll Llong beds represent the top of the Formation, reflecting a 'transitional' sequence of thin shelf sandstones which pass with into overlying mudstones of the Penmaen Dewi Formation (Table 3.2.).

5. Trwyn Llwyd Formation (D) – *(Whitlandian Stage)*

The term Trwyn Llwyd Formation is introduced for a complex siliciclastic-argillic shelf sequence which dominates much of the geology between Abercastle and Porth Maenmelyn (Map 3.). The Formation is characterised by systematically ordered facies associations; although nowhere is it exposed in it's entirety, having been pieced together from fault bounded successions at several localities between Pwll Deri and Pwllstrodur (Map 3.).

Based on the recognised facies associations, division of the Formation whilst not straightforward due to diachronaity, is comparatively well appreciated and grouped here into 5 Members referred to as the Pen Deudraeth Member (prodelta/shelf mudstones), Aberbach Member (deltaic sandstones/distributary mouth), Godor Mabli Member (heterolithic inner shelf clastics), Ogof Felin Member (shoaling bar/barrier sandstones), and Ogof Aderyn Members (heterolithic shelf clastics). The ability to identify this sequence throughout the coastal districts, in particular the Aberbach Member which even when highly deformed due to its conspicuous sheet sandstones acts as a marker horizon (Plate 3.5. a&b), has greatly aided the stratigraphic interpretation of the area, whilst alleviating some of the incongruent structural relationships. The Formation is argued to be of Whitlandian age {4.4.} and the probable diachronous equivalent of the Abercastle Formation; which together, may reflect the development of a shelf margin clastic wedge preceding the onset of deep water conditions from mid-Arenig times.

***Pen Deudraeth Member** (D1):* The Pen Deudraeth Member represents a fissile black marine shale sequence which forms the base of the Trwyn Llwyd Formation. The shales are in character with younger shales (i.e. Arenig – Llanvirn Shales; Table 3.1.), a distinction between which can only be made in the light of the Member's association with the overlying sandstones of the Aberbach Member {4.4.}.

***Aberbach Member** (D2):* The transition from the Pen Deudraeth Member into the Aberbach Member is marked by a gradational increase in both grains size and lateral continuity of bedding, reflecting a shallowing of marine conditions. The Aberbach Member represents a series of unfossiliferous sandstones, siltstones and mudstones characterised by coarsening-up cycles, the thickness of which is seen to vary, whilst internally the nature of inter-cycle sediments is often complex; departure from 'classic' mineralogical cycles being the norm. However, all are characterised by laterally persistent sheet sandstones at the top of individual cyclic units, making this Member, widely recognisable in coastal section (Plate 3.5.a&b). Together the Pen Deudraeth Member and Aberbach Member are interpreted as a prodelta mudstone and delta-front sandstone sequence {4.4.}.

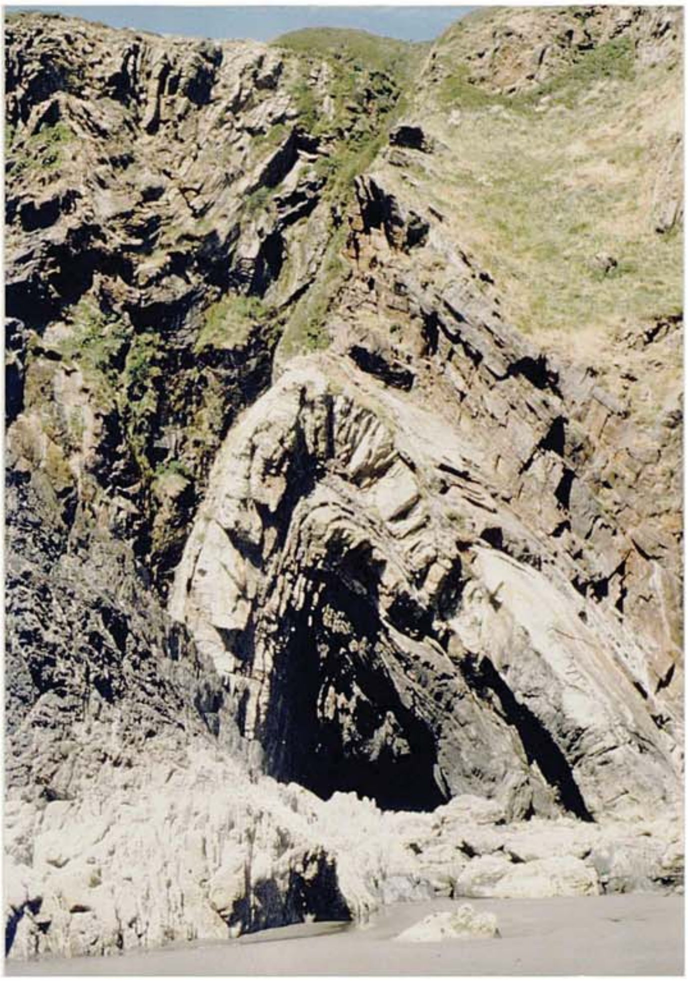
***Godor Mabli Member** (D3):* The Aberbach Member passes gradationally into a thick (≈160m) sequence of heterolithic sandstones and mudstones termed here the Godor Mabli Member. The Member is invariably deformed (Plate 3.3. and Plate 3.5), although appears complete within the cliff section west Godor Mabli to the headland of Trwyn Llwyd [S.M. 8826 3656 to S.M. 8896 3664]; serving as an excellent type-section as it illustrates passage from the underlying Aberbach Member and the transition into overlying Ogof Felin Member. The Member is characterised by alternating, laterally continuous (>300m) thin to medium bedded, fine- to medium sandstones and bioturbated mudstones/siltstones of the *Cruziana* association. The Member shows features typical of inner siliciclastic shelf sediments, where processes were operating at two different intensities; principally during storm and fair-weather conditions.

***Ogof Felin Member** (D4):* The Ogof Felin Member represents a massive to thick bedded, medium- to coarse sandstone and granulestone facies that separates the Godor Mabli

Plate 3.5. A & B. Comparative field photographs of the development of the Aberbach Member at Pwllchrochan [S.M. 8845 3660] and west of Godor Mabli [S.M. 8820 3670] typifying the incongruous nature of the structure, although clearly showing even where highly deformed the conspicuous sheet sandstones of the Member act as marker horizon (See also Plate 4.7. of the Aberbach Member developed on the headland at Abermawr 1km to the south).

Plate A. Shows the effects of bedding thickness and competency contrast in defining the nature and styles of folds in the area. The disharmonic folds in the top right corner are developed is the Godor Mabli Member which at this point can be seen to be relatively open although pass eastwards along the coast for 200m to where Plate B is taken and the folding in the same section approximates upright chevron folds below a major thrust surface. The Aberbach Member in comparison is weakly cleaved. In **Plate B.** both the Aberbach and the Godor Mabli are at the same level due to a minor cross-fault, the Aberbach Member at this locality rests above the folded Godor Mabli Member directly around the corner from where the figure is sitting. Note the faint increase in silt and fine sand lamination from the lower sheet sandstones, through a small and thin 'storm' sandstone passing abruptly into the sheet sandstone in top right hand corner. (n.b. the soft-sediments fold in Plate 3.1. occur directly above this locality, where the Aberbach Member passes gradationally into the Godor Mabli Member which shows no evidence of deformation).

A



B

Member from the overlying Ogof Aderyn Member. The Member is atypical due to its thickness (> 40m) and textural maturity interpreted to reflect temporary shoaling above fair weather base analogous to an offshore bar {4.4.}; its maturity makes the succession readily identifiable at outcrop and aids the interpretation of stratigraphic repetition.

Ogof Aderyn Member *(D5):* The term Ogof Aderyn Member is introduced here for a succession of heterolithic shallow marine clastics and mudstones that overlay conformably the Ogof Felin Member to the north of the Trwyn Llwyd headland [S.M. 8795 3670]. The base of the Member is represented by wavy and lenticular bedded mudstones and siltstones, amalgamated within coarse sandstones and granulestones of the Ogof Felin Member; although access to the Member is limited, precluding detailed interpretation. Lithologies do however, show striking similarities to the Godor Mabli Member, although this is not unexpected as the intervening Ogof Felin Member is interpreted to represent a transient clastic facies, with the Ogof Aderyn Member reflecting a return to siliciclastic-argillic shelf sedimentation.

6. Arenig-Llanvirn Shales (F) – *(upper Middle Arenig to lower Upper Llanvirn)*

Deposition of arenaceous material appears to have abated during the late middle Arenig (i.e. upper Whitlandian) from which time graptolitic marine mudstones accumulated in a relatively deep marine environment, conditions that were to persist until the late Upper Llanvirn. Subdivision of the mudstones (Table 3.1., i.e. Penmaen Dewi Formation (F1), *D. artus* Shales (F2), *D. Murchisoni* Shales (F3)) is based on faunal grounds {2.2.}, in the absence of which they are collectively referred to here as the Arenig-Llanvirn Shales. During this period of essentially 'stagnant' marine conditions subaqueous volcanism became widespread, evidence for which is exhibited principally by the Fishguard Volcanic Complex (see below), although 'isolated' volcanogenic mass-flows and subaqueous silicic pyroclastics with no recognised local source are a feature at all biostratigraphic levels. Other than igneous incursions, there appears no evidence for any major brake or hiatus, suggesting continuous mudstone deposition; although the successions have yet to provide the critical faunas to substantiate this inference. The environment of deposition is envisaged as partially restricted, within which sediment accumulation was mainly from suspension and fine-grained turbidity currents.

7. Fishguard Bay Group (FBG) – *Late Upper Llanvirn to Upper Llandeilo*

The term Fishguard Bay Group is introduced here for a thick succession (>1500m) of sandstones and mudstones believed to range in age from the late Upper Llanvirn to the Upper Llandeilo/basal Caradoc, occupying the low lying area between Fishguard and Goodwick (Map 1.) and the coastal districts from Fishguard to Dinas Head. The Group overlies the Fishguard Volcanic Complex with apparent disconformity {4.7.}, finning upwards from a lower arenaceous succession, referred to here as the Lower Town Formation, into thick marine mudstones from which rare fauna allow biostratigraphic division into the Dyffryn Formation (Lower Llandeilo) and *N. gracilis* Shales (Upper Llandeilo – Lower Caradoc). The *N. gracilis* Shales (the 'Llandeilo Shales' of Lowman & Bloxam 1981) crop out between Fishguard and Dinas Head east of this study area, although are regarded as part of the same depositional unit and appropriately discussed here in the context of a major stratigraphic sequence.

Lower Town Formation (G): The term Lower Town Formation is introduced for a succession of interbedded coarse sandstones, siltstones, mudstones, and minor volcano-clastics, that crop out on the headlands of Saddle Point [S.M. 9580 3783] and Castle Point [S.M. 9580 3775] at the harbour entrance of Lower Town, Fishguard. On the basis of lithologic variation the Formation has been divided into two members, the ***Castle Point Member*** (G1) and ***Pantycelyn Member*** (G2). The Castle Point Member relates to a lower succession of massive bedded coarse sandstones and subordinate mudstones of varying affinity. The typical lithology of the Pantycelyn Member is one of thin to medium planar bedded, sheet sandstones, polymictic conglomerates, and crystal-lithic tuffs, interbedded with black pyritiferous mudstones and siltstones; the mudstone-sandstone ratio is much higher than the underlying Castle Point Member, and scale of bedding finer. Both members are thought to reflect relatively deep marine accumulations, mudstones reflecting prolonged periods of predominantly 'background' suspension sedimentation, onto which other lithologies were superimposed as mass-flow deposits; the detrital mineralogy of which suggests a shallow marine origin with resedimentation into deeper water {4.7.}.

Dyffryn Formation (H) *(Lower Llandeilo):* The term Dyffryn Formation is introduced here for a thick (>500m), although poorly exposed succession of predominantly argillic sediments which overlie the Lower Town Formation (Map 1.). Based on varying lithology and graptolite fauna, the Formation falls naturally into two members referred to here ***Drim Member*** (H1) and ***Tref-wrgi Member*** (H2). The Drim Member is characterised by thin bedded calcareous mudstones. The typical lithology of the Tref-wrgi Member is one of highly cleaved black pyritiferous mudstones within which prolific graptolite horizons locally define bedding. Both members have yielded a fauna indicative of the *G. teretiusculus* Biozone, suggesting that a Lower Llandeilo age is appropriate.

***N. gracilis* Shales** *(Upper Llandeilo – Lower Caradoc):* Outside of this study area, occupying much of the coastal district between Penrhynychen [S.M. 9830 3840] and Dinas Head [S.M. 0050 4050], black mudstones referred to as the 'Llandeilo Shales' by Lowman & Bloxam (1981) have yielded a fauna which suggests the presence of the *N. gracilis* Biozone (i.e. Upper Llandeilo – Constonian Stage). There appears little doubt that the succession occupies a biostratigraphic position above the Dyffryn Formation whilst their lower beds may represent the along strike continuum of the Tref-wrgi Member {4.6.}.

3.4.2. PRINCIPLE IGNEOUS DIVISIONS

The existing igneous stratigraphy in the Fishguard – Porth-gain district is modified in the light of detailed mapping. The Fishguard Volcanic Complex represents the principle igneous succession, although the identification of successions which are assigned here to the late Arenig are new additions. Thin tuffaceous horizons are common to the Arenig-Llanvirn Shales at all biostratigraphic levels {5.2.}, although a source (or sources) is lacking; it is possible that small localised volcanic centres lie hidden within the inland sequence.

1. Arenig igneous activity *(Yns Castell Ash & Pwll Deri Tuffs)*

Two successions in the Fishguard – Porth-gain district, the Yns Castell Ash and Pwll Deri Tuffs (Table 3.1.), are thought here to reflect extrusive igneous activity during late Arenig times. Both successions are silicic in character and subaqueously emplaced, whilst appearing to occupy a similar stratigraphic position; it being possible that they represent part of the same event, or episode of events.

Yns-Castell Ash (YCA): The Yns-Castell Ash (*c.f.* Cox 1930), a bedded sequence of subaqueously emplaced crystal-lithic lapilli tuffs {5.1.}, are best observed on the headland of Yns y Castell at the entrance of Abercastle Harbour [S.M. 8510 3385], where they are seen to lie with concordance upon shales regarded here as belonging to the Penmaen Dewi Formation (i.e. Arenig shales). The Tuffs form thick planar bedded units (*c.* 50cm to 5m), and are thought to reflect the products of 'proximal' mass-gravity flows derived from a volcanic source in the immediate vicinity of the St. Davids – Fishguard – Prescelly district {5.2.}.

Pwll Deri Tuffs (PDT): The Pwll Deri Tuffs is a collective term for a newly recognised sequence of subaqueous pyroclastic flows and volcanoclastics. The *Tuffs* which are very poorly exposed, crop out discontinuously in the vicinity of Pwll Deri [S.M. 8915 3825], interbedded with shales from which Cox (1930) records the presence of extensiform graptolites, strongly suggesting that an Arenig age is appropriate. The variety of lithologies observed include; welded ash-flow tuffs, crystal-lithic tuffs, and various silicic breccias. The presence of welded tuffs indicate that true pyroclastic lavas constitute part of the sequence; although, whilst there is little doubt that the sequence was emplaced within the marine environment, whether or not they represent the product of subaqueous or subaerial eruption is uncertain {5.3.}

Arenig intrusive activity: The majority of intrusives within the Fishguard – Porth-gain district are of Llanvirn age related in part with the development of the Fishguard Volcanic Complex and prevailing volcanotectonic environment. However, whilst it is commonly assumed that intrusive activity throughout SW Dyfed is Llanvirn age, an assessment of magma/sediment relationships of auto-brecciated felsic sills within Penmaen Dewi Formation (i.e. Arenig shales), indicate that emplacement and sedimentation were penecontemporanueous suggesting that an Arenig age is appropriate {5.3.}; it being probable that specific intrusive suites reflect the hypabyssal counterparts of Arenig extrusives.

2. Fishguard Volcanic Complex – *Llanvirn Igneous activity*

The Fishguard Volcanic Complex is used here as a collective term to incorporate all associated volcanic and intrusive successions which crop out in a continuous E-W trending belt (>25 km), from Porth Maenmelyn in the west to the Prescelly Hills in the east (Fig. 2.2.). The development of the Volcanic Complex east of Fishguard has been discussed in detail by Lowman (1977) and Lowman & Bloxam (1981). They describe a variety of silicic lithologies, including coherent lavas, fall-deposits, and epiclastics, although suggest that the majority of rock types are welded pyroclastic flows, which originated from a terrestrial source and emplaced within a subaqueous environment possibly to depths as great as 1.5 km. As outlined {1.3.}, the development of the Volcanic Complex west of Fishguard (this study area) is dominated by thick basaltic and silicic volcanics of various affinity, all of which were erupted within the subaqueous environment, with little or no evidence for the products of subaerial eruption or emplacement of flows from subaerial vents. Welded pyroclastic flows appear to be absent. The Volcanic Complex requires further investigation with an attempt to 'blend' the compositional, lithological and environmental variations that appear to exist between the successions east and west of Fishguard. The terminology used by previous workers in describing the various successions is somewhat confusing. For example, Lowman & Bloxam (1981) state than the Fishguard Volcanic Group is almost entirely rhyolitic in composition. However, to the west of Fishguard the thickest development of basaltic lavas in Wales crop out on the Pen Caer peninsular, similarly referred to as the Fishguard Volcanic Group (Bevins & Roach 1979a., 1982; see below). It is not attempted here to revise the terminology, this requiring regional mapping through the entire strike length of the Volcanic Complex.

Fishguard Volcanic Group

The term Fishguard Volcanic Group was introduced by Bevins & Roach (1979b) to incorporate the spatially observed tripartite volcanic succession west of Fishguard (i.e. Porth Maenmelyn Volcanic Formation, Strumble Head Volcanic Formation, and Goodwick Harbour Volcanic Formation (Fig. 1.4.; Table 3.1.)). Such terminology is retained here, in part avoiding unnecessary interference with existing nomenclature, although principally to aid communication rather than the representation of discrete volcanostratigraphic units. There are too many contradictions within the stratigraphy to suggest its ardent use, although to substitute the subdivision would at present unnecessarily complicate the sequence. It should be noted however, that the picture portrayed by previous workers, in their apparent ability to trace a conformable tripartite volcanic sequence from Porth Maenmelyn to Fishguard can not be verified (Maps 1-2). Nowhere can such division be identified at outcrop south of Pen-rhiw [S.M. 9420 3900], whilst the tripartite relationship on the Pen Caer Peninsula appears to be more apparent than real. Irrespective of stratigraphic attitude however, all silicic volcanics underlying and overlying the basic volcanics of the Strumble Head Volcanic Formation are referred to as the Porth Maenmelyn Volcanic Formation and Goodwick Volcanic Formation respectively.

Porth Maenmelyn Volcanic Formation (PMVF): The Porth Maenmelyn Volcanic Formation represents a thick accumulation of rhyolitic clastics, coherent obsidian flows and hypabyssal intrusives (Maps 1-3 & 5.). The lithotype section for the Formation crops out within the cliffs on the north side of Porth Maenmelyn; the succession having recently been documented in detail by Bevins & Roach (1982) and Kokelaar *et al.* (1984b). Map (5) is a map of the coastline between Pwll Deri and Pen Bush [S.M. 8805 3965] which incorporates the type section as described by previous workers; although also documents the Formation's development to the south at Aber Twn [S.M. 8890 3865] showing the highly fragmented nature of the volcanostratigraphy and repeat successions.

Strumble Head Volcanic Formation (SHVF); The Strumble Head Volcanic Formation spatially occupies the majority of the Pen Caer Peninsula (Maps 1-2), and is composed of a variety of subaqueous basaltic products. Basaltic volcanism was generally effusive in nature, pillowed lavas and tube networks volumetrically dominate; however, massive flows, pillow breccias, high level contemporaneous intrusives, bedded basaltic hyaloclastites and basic volcanoclastics developed locally {5.4.}. The deposits represent a complex interplay of extrusive, hydroclastic, autoclastic and gravity driven processes, within the subaqueous environment {5.4.}. Inland exposure on the Pen Caer Peninsular is extremely poor, although coastal exposures between Porth Maenmelyn - Strumble Head (perpendicular to strike) and Strumble Head to Maen Jaspis (parallel to regional strike) allow an excellent insight into the various products of subaqueous basaltic volcanism. The term ***Carreg Gybi Member (CGM)*** is introduced here for a thick (≈ 30m) succession of interbedded volcanoclastic turbidites, pyroclastic debris flows, marine shales, and porcellanitic mudstones (Map 6), that occupy a position towards the top of the Strumble Head Volcanic Formation. The Member is one of the few discrete lithologic units present within the Volcanic Formation, whilst it is of interest as it allows an insight into the processes that were operative during periods of prolonged quiescence from active volcanism.

Goodwick Volcanic Formation (GVF): The Goodwick Volcanic Formation crops out extensively in the vicinity of Maen Jaspis and Y Penryhn [S.M. 9450 4050]. The Formation is dominated by rhyolite breccias, flow banded obsidian lavas and crystal tuffs which can be viewed as in general character with the Porth Maenmelyn Volcanic Formation. The Formation exhibits a complex interplay and association of coherent flows, *in situ* auto-breccias, and gravity driven mass-flow deposits, all emplaced within a subaqueous environment. The Formation has recently been discussed at length by Kokelaar *et al.* (1984b), details of which are given in Chapter 5.

Contemporaneous Intrusives: Contemporaneous intrusive activity associated with the Fishguard Volcanic Group can be demonstrated to have been widely variable regarding gross habit and associated emplacement mechanisms {5.4.}. Their compositional range has been shown to range from basic to intermediate (Bevins 1979, 1982), with the development of gabbros and tonalites, and finer grained equivalents.

3.5. Summary

The stratigraphy and the structure depicted on Maps 1-6, appears at first over complex, although such complexity can be viewed as consistent with the earlier findings of Cox (1916, 1930). The recognition of large scale repetition has in part resolved some of the stratigraphic re ationships, details of which are summarised in subsequent chapters.

It is thought that the principle control on the regional distribution of the geology is best explained in terms of two phases of Caledonide deformation, the first has an E-W trending old-thrust signature, the second relates to NE–SW cross-faults. The E-W structure whilst not typically Caledonide, is believed to be the principle fabric and associated with the progressive swing of major structures from Central to SW Wales; possibly reflecting basin margin parallelism. As inferred by Fitches & Cambell (1987) there is almost a preoccupation with the NE–SW trend; this appears to be verified in the Fishguard - Porth-gain district, in as much as the NE–SW structure whilst present have limited control on the geology.

An extension of the fold-thrust trend east towards Carmarthen may be likely in as much as Cope (1979) documents similar stratigraphic complexity resulting from southerly direct thrusts, associated with shallow cover-basement relationships. It can be speculated that this style of deformation dominates the regional geology of SW Dyfed possibly in a regional fold-thrust belt, although is obscured by subsequent cross-faults. Evidence to support this suggestion comes notably from major structures to the south of Fishguard where folds overturn and verge to the south (i.e. Abereiddi Bay Syncline (Waltham 1971), St. Davids Syncline (Roach 1969), Llanrian Syncline (Cox 1930) and Tremaenhir Syncline (Williams

1934)), all of which show evidence of contractional faults (thrusts), which either cut-up through the fold structure or are in close association with the inverted limb; suggesting that they are geometrically related allowing accommodation and further shortening. Whilst structural data is limited, the overall tectonic style throughout the region appears to be akin to 'thin-skinned' tectonics (see Winslow 1981;Figure 6.).

There can be little doubt from the observations of fault-rock products and associated polymineralic assemblages that deformation within the area allowed the migration of metamorphic fluids. However, the question may be asked is metamorphism a function of deformation (i.e. strain-heating), or did deformation, facilitate mineral growth merely by creating conduits through which fluids migrated, and so reflect a pre-existing geothermal gradient. The question can not be answered from mineral chemistry alone (Chapter 6) and no conclusions can be drawn other than stating that at the time of deformation conditions of metamorphism were of the greenschist and prehnite-pumpellyite facies whilst phase development was pre-, syn- and post deformation. However, this suggests that metmorphism in the Welsh Basin can not be categorised into discrete metamorphic regimes and each area must be taken on its own merit.

CHAPTER FOUR
SEDIMENTARY STRATIGRAPHY AND SEDIMENTATION

4.1. INTRODUCTION

The sedimentary sequence exposed within the Fishguard – Porth-gain district is believed to represent a fragmented insight into the geological development of SW Dyfed for approximately 100 Ma. (i.e. Middle Cambrian to Upper Llandeilo). During this time the nature of sedimentation varied considerably, reflecting localised change within the depositional system and the influence of larger, regional eustatic events. The aim of this chapter is to evaluate such change in proposed chronologic order, by describing the stratigraphy, age constraints and sedimentary characteristics of all newly introduced stratigraphic units.

The sedimentary sequence is discussed under broad stratigraphic headings (Table 4.1.), which for the most part corresponds with the major lithologic changes. In the general absence of faunal evidence, age constraints are placed through likely inference and conjectural correlation with strata from elsewhere in SW Dyfed. However, regional correlation is not sufficiently sensitive and the sedimentology not sufficiently detailed, to address aspects of SW Dyfed palaeogeography other than in passing.

4.2. MIDDLE CAMBRIAN (?) STRATIGRAPHY AND SEDIMENTATION.

4.2.1. Lach Dafad Formation (A). *Cumulative Thickness (> 120m).*

The Lach Dafad Formation is a collective term for distinctive red-green sandstone-mudstone successions that crop out at numerous localities between the coastal districts of Pwllstrodur and Penbwchdy (Maps 3 & 4.). Division into the Carreg Herefio, Tregwent and Pwlldawnau Members (Table 4.1.) is based on varying lithology. Each member is fault bounded, although in the absence of biostratigraphic evidence it is realistic to assume that they represent fragmented members of the same lithostratigraphic unit; the spatial attitudes of which is made by conjecture (Fig. 4.1.). The Tregwent Member is very poorly exposed and precludes interpretation.

Age constraints: By analogy with strata elsewhere in SW Dyfed, the red-green variegated colouration of the various sediments which constitute the Lach Dafad Formation, strongly suggests that the Formation occupies a position within the lower part of the Cambrian. In the absence of faunal evidence the Formation can arguably be correlated with sediments of both Lower and Middle Cambrian age {2.2}. From the author's personal observations and documentation of the Cambrian 'red-beds' of SW Dyfed (e.g. Cox *et al.* 1930a; Turner 1979;

Table 4.1. Stratigraphic subdivision of the sedimentary sequence developed throughout the Fishguard – Porth-gain district. For distribution of individual stratigraphic units, see Maps 1-4.

SYSTEMS, SERIES, STAGES & BIOZONE			STRATIGRAPHY			Thickness		Lithology	Inferred depositional environment	Correlated successions in S.W. Dyfed referred to in the text
ORDOVICIAN	LLANDEILO	UPPER *N. gracilis* Biozone	Fishguard Bay Group	*N. gracilis* Shales (Hendre Shales)			> 1000m	Graptolitic marine shales see Lowman (1977) and Lowman & Bloxam (1981)	Marine mudstones restrict and possibly deep marine environment	Hendre Shales (Carmarthen) Llandeilo/Caradoc Shales (Abereiddi Bay)
		LOWER *G.teretiusculus* Biozone		Dyffryn Formation	Tref-wrgi Member	> 500m	> 200m	Graptolitic marine shales; diffuse fine silt grade lamination, deposition mainly from suspension and pelagic processes.		G. teretiusculus Shales (Carmarthen District)
					Drim Member		> 300m	Graptolitic calcerous marine mudstones		
				Lower Town Formation	Pantycelyn Mb.	> 220m		Marine shales interbedded with thinly bedded coarse sandstones and polymictic conglomerates deposited by density modified grain flows and debris flows.	Deep Marine (althought shallow water conditions prevailing locally)	Fairfach Group (Carmarthen/Llandeilo)
	LLANVIRN				Castle Point Mb.			Massively bedded coarse sandstones and conglomerates, occassional fine marine mudstones; deposition by mass-flow currents		
		UPPER *D. murchisoni* Biozone	Arenig – Llanvirn Shales	*D. murchisoni* Shales			> 150m	Arenig – Llanvirn Shale Black graptolitic marine shales. Deposition, in the main from suspension pelagic processes and fine grained muddy turbidites. Fine 'isolated volcanoclastic sediments	Restricted basin mudstones variable depth	
		LOWER *D. artus* Biozone		*Artus* Shales			>100m (?)			
	ARENIG	FENNIAN		Penmaen Dewi Formation			> 120m			
		WHITLANDIAN	Abercastle Formation	Pwll Llong beds		> 90m	c. 20m	Fine sandstone/mudstone alternations passing conformably into the Penmaen Dewi Fm.	Thin shelf-sandstone to shelf mudstones	
				Pen Porth-egr member			> 50m	Fining-up transgressive sedimentation: coarse tabular cross bedded sandstones	Shore-face to inner shelf ?	
			Trwyn Llwyd Formation	Carreg Gwynion Mb.		> 350m	> 75m	Heterolithic sandstone – mudstone intercalations, storm and wave structures	Inner Shelf to ? Outer Shelf	Abercastle Formation
				Ogof Felin Mb.			c. 45m	Large scale tabular cross-bedded coarse sandstones	Emergent offshore bar complex	
				Godor Mabli Mb.			c. 150	Hetrolithic sandstone – mudstone intercalations, *Cruziana* association	Inner shelf	
				Aberbach Mb.			50 65m	Coarsening-up cyclic sedimentation; mature sheet sandstones	Delta front sandstones	
				Pen Deudraeth Mb.			> 45m	Black marine mudstones, with silt-grade turbidites and fine storm layer sandstones	Shelf Mudstones: Prodelta sequence	
		MORIDUNIAN	Porth-gain Harbour Formation	Shale member		> 40m	> 20m	Black shales, with fine discontinuous silt and fine sandstone	Transgressive marine sequence tidal sandstones passing into subtidal mudstones	Ogof Hên Formation (Ramsey Island, Carmarthen)
				Sandstone member			c. 20m	Polymictic conglomerates, passing into hetrolithic wavey, flaser, and lenticular bedded mudstone sandstone alternation		
CAMBRIAN	UPPER (Merioneth Series)		*Lingula* Group	Porth Ffynnon Formation		> 200m	> 100m	Heterolithic fine sandstones and siltstones, wave and current generated structures	Shallow Inner Marine Shelf	'Lingula Flags' (St. Davids, Prescelly Hills, Carmarthen)
				Mynydd Morfa Formation			> 200m			
	MIDDLE (St. Davids)		Lach Dafad Formation	Pwlldawnau Member			> 70m	Red-green massive planar bedded sandstones fining-up into red laminated mudstones	Shallow marine tidal and wave dominated	Solva Group (St. Davids District)
				Tregwent Member			> 55m	Red mudstones with diffuse silt-grade parallel lamination, passing into heterolithic alternating fine sandstones and siltstones. Evidence for wave generated structures		
				Carreg Herefio Mb.			> 40m	Large tabular and trough cross-bedded coarse sandstones, interbedded with thin wavy and lenticular mudstone - sandstone alternations; frequent herring bone cross lamination		
	LOWER (?Comley Series)									

Williams & Stead 1982), it is thought that the Formation's closest lithostratigraphic parallel are sediments of the Lower Middle Cambrian Solva Group (Table 4.2.); showing marked similarities with the development of the Group between Porth Clais [S.M. 7425 2340] and Ogof Lle-sugn [S.M. 7360 2335, see I.G.S. Special Geological Sheet, St. Davids: SM 72 & Parts of SM 62, 73; 1:25000]. The grey sandstones of the Carreg Herefio Member may correlate with the Grey Sandstone Member of the Middle Solva Beds, a lithological unit which appears to be evident throughout much of SW Dyfed (Williams & Stead 1982); whilst the Tregwent and Pwlldawnau Members show similarities with the overlying Purple Sandstone Member (Table 4.2.).

The nearest recorded fossiliferous sediments of the Menevian Group (i.e. Upper Middle Cambrian), are those documented from the Tancresdton area to the south-east of Fishguard. In this area Williams (1934) records the presence of a variety of fauna, including *Eodiscus* cf. *punctatus* Salter, a fauna generally restricted to the *Hypagnostus parvifrons* and *Ptychagnostus punctuosus* biozones (Thomas *et al.* 1984), broadly equivalent to the previously held *P. hicksii* Biozone; characteristic of the lowermost Menevian Beds in SW Dyfed (Stubblefield 1956). Williams (1934), indicates that the Menevian Group in this area is greater than 90m thick, although the sediments are generally nondescript, composed of homogenous blue-green pyritiferous mudstones; there is little evidence to indicate the existence of similar strata in the Fishguard – Porth-gain district.

(a). Carreg Herefio Member (A1). *Type Locality (S.M. 8825 3596) Thickness (>35m).*

The Carreg Herefio Member is best studied at the wave-cut platform on the headland of Lach Dafad [S.M. 8810 3578]. The Member is dominated by a lower succession of buff-white to dull grey-green fine- to coarse sandstones and subordinate mudstones (*Grey Sandstone association* Fig. 4.1.), overlain by coarse red-green sandstone and red shales (*Red-bed association* Fig 4.1.). The transition between the two variants appears conformable, marked by a gradational variegated change in colour, and change in bedding from massive to discrete planar units. The thickness of the Member at outcrop is in excess of 35m, although its base is fault bounded and its upper beds truncated by a coarse gabbro (Map 3.).

Description: Texturally mature *tabular cross-bedded sandstones* dominate the accessible exposure, consisting of thick to massive bedded units of coarse grey sandstones and subordinate mudstones (<5%). Individual co-sets are generally planar, varying in thickness from 40cm to 1.5m, and are frequently defined by thin (1-2cm) laterally persistent wavy and lenticular mudstone partings. No bioturbation is present within the intervening mudstones, which are commonly reworked; evidence for which is seen by rounded mudstone chips lining the amalgamation surface between mudstone and sandstone. Internally co-sets are dominated by thick tabular cross-sets with relatively high-angle discordances. Foreset lamination is commonly graded from granulestone/coarse– to medium/fine sandstone giving the appearance of a couplet. Small scale trough cross beds may occur at the base of individual co-sets. The sandstones are compositionally immature, composed of quartz, volcanic rock fragments, sedimentary rock fragments, strained polygonal quartz aggregates, feldspar, and minor intraformational mudstones.

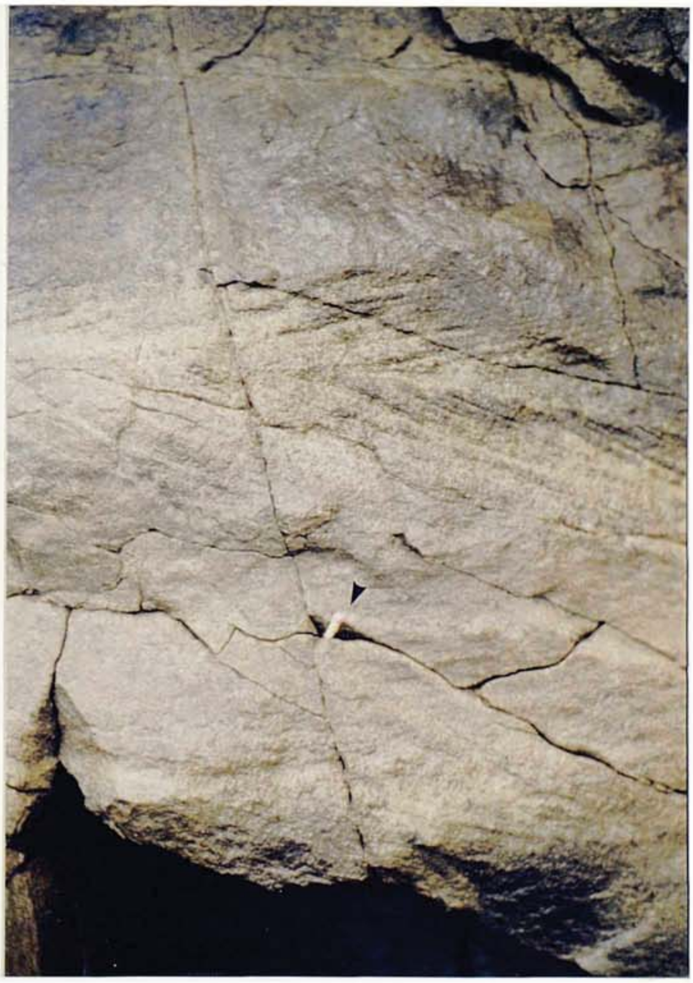
A texturally immature variant on the tabular cross-bedded sandstones are thick moderately sorted *trough and tabular cross-bedded* sandstones (Plate 4.1.) readily discriminated by a deeper weathering profile and grey green colouration. Both immature and mature lithologies are amalgamated on a metre scale with little or no systematic order. Mudstone drapes are common along reactivation surfaces within tabular sets. The most distinctive feature of the sequence is the wide spread presence of herringbone cross-lamination, reflecting current reversal (Plate 4.2.). Bioturbated *mudstones* occur at sporadically forming deeply weathered horizons which vary in thickness from several centimetres to tens of centimetres. Within the mudstones structures are limited to weak wavy and flaser bedded mudstone–siltstone alternations and rare fine-coarse sandstone to granulestone lamination.

Planar bedded red-green sandstones and red mudstones occupy the core of the Treseisyllt syncline, spatially overlying the grey sandstones on the wavecut platform at Lach Dafad. The succession concomitantly fines-up, from thick bedded coarse to medium sandstones with occasional red and green mudstone partings, into fine laminated buff-red mudstones with rare fine sandstones. Bedding is however, generally diffuse due to extensive diagenetic alteration with abundant nodules forming bedding planar horizons and obscuring primary structures. Nodules are variable in size (5-30cm), commonly containing small amounts of copper mineralisation (< 1%); malachite being readily identifiable in hand specimen, whilst traces of chalcopyrite, covellite, bornite and chalcocite have been observed in polished thin-section. Where sandstone beds are sufficiently preserved, they show lateral continuity through outcrop, within which the most common sedimentary structure is planar lamination, although individual beds commonly display extensive penecontemporaneous deformation structures (Plate 4.3.).

Interpretation: The lower grey sandstones reflect deposition in an shallow marine environment influenced by tidal and wave activity. The *texturally mature tabular cross-bedded sandstones* may reflect deposition within the marine shoreface, with thin mudstone veneers commonly reworked by sand wave migration representing prolonged periods of 'slack-water' sedimentation. From the frequent occurrence of reactivation surfaces and the presence of herringbone cross lamination, the *immature trough and tabular cross-bedded sandstones* are

Plate 4.1. Trough cross-bedded texturally and compositionally immature sandstones from the Carreg Herefio Member [S.M. 8825 3596).

Plate 4.2. Herringbone cross-lamination with the Carreg Herefio Member [S.M. 8825 3598]. Such sedimentary structures are common to the lower parts of the Member, suggesting current reversal with a tidal influenced environment. The scale of the photograph is indicated by a match head which is arrowed



thought to reflect less vigorous conditions under tidal influence, and are true 'tidalites' (*c.f.* Klein 1971); although channel and wave migration occurred episodically as indicated by the intercalation of thick flazer and wavy bedded partings. The coarse grain size of the sediments and compositional and textural immaturity are analogous to clastic intertidal and barrier sequences (e.g. Carter 1978).

The *planar bedded red-green sandstones* are not readily discriminated. The predominance of planar lamination may reflect rapid deposition as high-energy storm generated sheet sandstones, such an origin may account for the extensive convolution of individual beds which probably resulted form the upward dispersion of trapped overpressured pore fluids. Such a sequence may represent proximal storm deposits (Johnson 1978), and record deepening of the marine environment from the underlying tidal sequence. The fining-up into red mudstones may reflect, either; further deepening to a point where deposition was below the effective limits of wave or storm sedimentation, or migration of the sandstone source. The presence of thin sandstones may suggest that the latter is not the case.

(b). Tregwent Member (A2). *Type Locality [S.M. 8945 3496], Thickness (>50m)*

The Tregwent Member represents a distinctive, yet poorly exposed succession of unfossiliferous marine red shales and thin bedded red-green sandstone and siltstones that crop out in the vicinity of Tregwent Wollen Mill [S.M. 8945 3496] and Treseisyllt Hill [S.M. 8915 3517]. The numerous small quarries near the entrance to New Mill Farm, from where Cox (1930 p.278) describes the lithology are unfortunately overgrown, although recent excavations at Tregwent Wollen Mill allows the salient lithological characteristics to be discerned. The exposed thickness of the Member is in excess of 55m. The base and top of the Member is nowhere seen, and appears to be totally fault bounded (Map 3). The lower red shales are thought to represent the equivalent of the red mudstones observed at the top of the Carreg Herefio Member (Fig. 4.2.).

The *red shales* are characterised by their homogenous fine grain size and red colouration, although generally nondescript. Reddening occurs throughout the succession, there appearing to be no evidence for any variegated colouration as is common to other 'red-bed' strata. The shales possess a strong penetrative cleavage, whilst locally thin silt-grade lamination enables a diffuse bedding to be defined. Lamination shows no evidence for grading or other depositional features. Evidence for bioturbation or other recognisable sedimentary structures are absent.

Thin medium bedded *heterolithic red-green sandstone-mudstone alternations* spatially overlie the red shales. Their exposed thickness is in excess of 25m, although they probably exceed this estimate. Outcrop is restricted to discontinuous exposure in the form of a small topographic ridge above the area occupied by the red shales in the fields north-east of Garn

Barcud [S.M. 8920 3505]. The lithology is seen to be dominated by thin interbedded siltstones and fine-medium sandstones. Colouration is variegated, although silt–grade horizons are generally grey-green, whilst coarser sandstone beds are commonly reddened. Fine mud-drapes on ripple surfaces are locally evident, which may suggest deposition within a relatively shallow marine environment.

(c). The Pwlldawnau Member (A3).*Type Locality S.M. [8800 3700]; Thickness (>75m).*

The Pwlldawnau Member crops out in the vicinity of Pwlldawnau [S.M. 8800 3700]. The Member shows a complex range of lithologies, although the environment (or environments) of deposition can not be readily discriminated. Beds which show a gross lithological similarity to those at Pwlldawnau, underlie the Carreg Herefio Member along cliffs at Pwll March [S.M. 8835 3612], whilst a further succession of red-green variegated sandstone and mudstones outcrop to the north of Pwllstrodur on the Mynydd Morfa headland [S.M. 8695 3440]; both successions are equated with the Lach Dafad Formation (Map 3.).

The Member comprises; laminated red mudstones, thin bedded siltstone/sandstones and channeled granulestones, and an associated volcanoclastic mass-flow (Fig 4.1.). The laminated red mudstones dominate the lower part of the Member. The thin bedded siltstone/sandstones and channeled granulestones dominates the upper part; onto which the volcanoclastic mass-flow was superimposed as a single depositional unit.

Description: The *laminated red mudstones* are characterised by buff red mudstone with fine planar laminated siltstone and rare fine sandstone. Lamination is commonly normal graded although no scouring appears evident. Bioturbation as first noted by Cox (1930) is locally prolific where lamination is present in sufficient quantity to highlight its presence, much of which resembles the large vertical burrows of *Skolithos* .

Thin bedded mudstone/sandstone and channelled granulestones dominate the upper part of the Member. The lithology is characterised by thin (1-5cm) bedded red and green laminated mudstones and fine to medium sandstones. Bedding contacts are generally planar, whilst internally individual beds may exhibit both planar lamination, and less frequently, fine cross-ripple lamination; the latter causing undulatory swell and pinch bedding surfaces. The cross laminae may show grading from fine sandstone to mudstone, the latter forming thin mudstone drapes (flazers). Sructureless fine red mudstones may occur in ripple troughs and less frequently drape subjacent ripples forming continuous mudrock bands. Internally fine lamination may show evidence for convolution. Throughout this alternating lithology, thin (<50cm) discrete channelled granulestones are common. Channeled bases are scoured, the erosive properties is attested to by the presence (in thin section) of angular 'rip-ups' from underlying lithologies. Upper bedding surfaces are generally planar, although locally show evidence for tectonic modification. The granulestones are framework supported and compositionally immature, containing a wide variety of volcanic (predominantly silicic) and sedimentary rock fragments, K-feldspar and strained polygonally quartz aggregates.

A distinctive deeply weathered tuffaceous lithology regarded as a *volcanoclastic mass-flow*, occurs towards the top of the Member (Plate 4.4.). The lithology, which remains enigmatic in its origin, appears to have been emplaced as a single depositional unit. The basal contact is planar to undulatory, whilst weakly erosive properties are suggested by the presence of 'rip-up' clasts from underlying lithologies. The lithology shows no obvious features of sorting and little evidence for any internal structure, being essentially a fine-grained crystal-lithic tuff; although the matrix is poorly resolved as a result of low-grade metamorphism and possible early diagenetic alteration. Within the tuffaceous matrix, bleached angular ash- to lapilli size fragments are readily recognised which may represent original vitroclasts. Isolated boulders of weakly porphyritic and vesicular lava whilst not common, are sufficiently numerous to support the contention that the deposit is volcanogenic in origin and has suffered little or no reworking (Plate 4.4.).

Interpretation: Sediments of the Pwlldawnau Member are not readily discriminated, due in part to the limited exposure and uncertainty of relations with successions elsewhere. The lowermost *laminated red mudstones* with *Skolithos* like burrows are almost certainly marine. However, the thin bedded *red-green sandstone/mudstone alternations – channelled granulestones* and *volcanoclastic mass-flow* must remain largely enigmatic, as there is insufficient width or access to exposure to allow the sequence to be evaluated in detail. A scenario may be that the succession represents transitional deposition between fluvial and marine environments (i.e. supratidal). Such speculation may not be too extreme, in so much as the preservation potential of a fine grained tuff, deposited as single event flow would be extremely low in a shallow marine environment; such a deposit would be expected to be rapidly reworked and dispersed. The origin of the flow remains a curiosity. However, if volcanoclastic, as is contended and the age constraints correct to the point where the Lach Dafad Formation is of the 'lower' Cambrian, it represents direct evidence for igneous activity in this part of Wales during the Cambrian.

Deposition within the Lach Dafad Formation

The Lach Dafad Formation represents a range of environmental conditions and spans a spectrum of processes. Sedimentary structures are sufficiently preserved within the Carreg

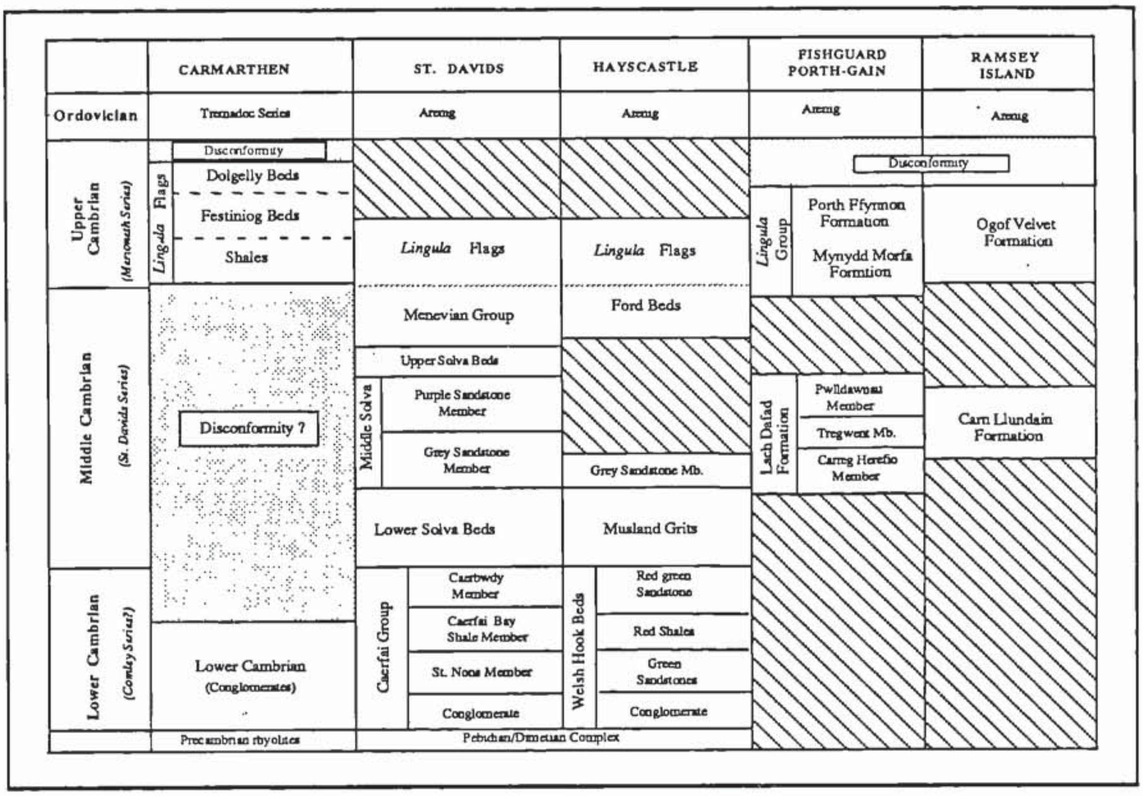
Plate 4.3. Extensively convoluted fine- to medium sandstones with the Carreg Herefio Member. Such penecontemporanueous structures may reflect the upward escape of trapped pore fluids in the underlying sediment. Note planar lamination of the sandstone with the 'dish' structure (arrowed).

Plate 4.4. *Volcanoclastic mass-flow.* Non-structured volcanoclastic mass-flow deposit within the Pwlldawnau Member [S.M. 8800 3700]. The deposit is thought to represent direct evidence for Cambrian volcanicity in this part of Wales. Note vesicular margin (V) on the edge of the weakly porphyritic volcanic boulder in center view, and fine tuffaceous lapilli-sized fragments (vitroclasts ?, TL), observed throughout the deposit.

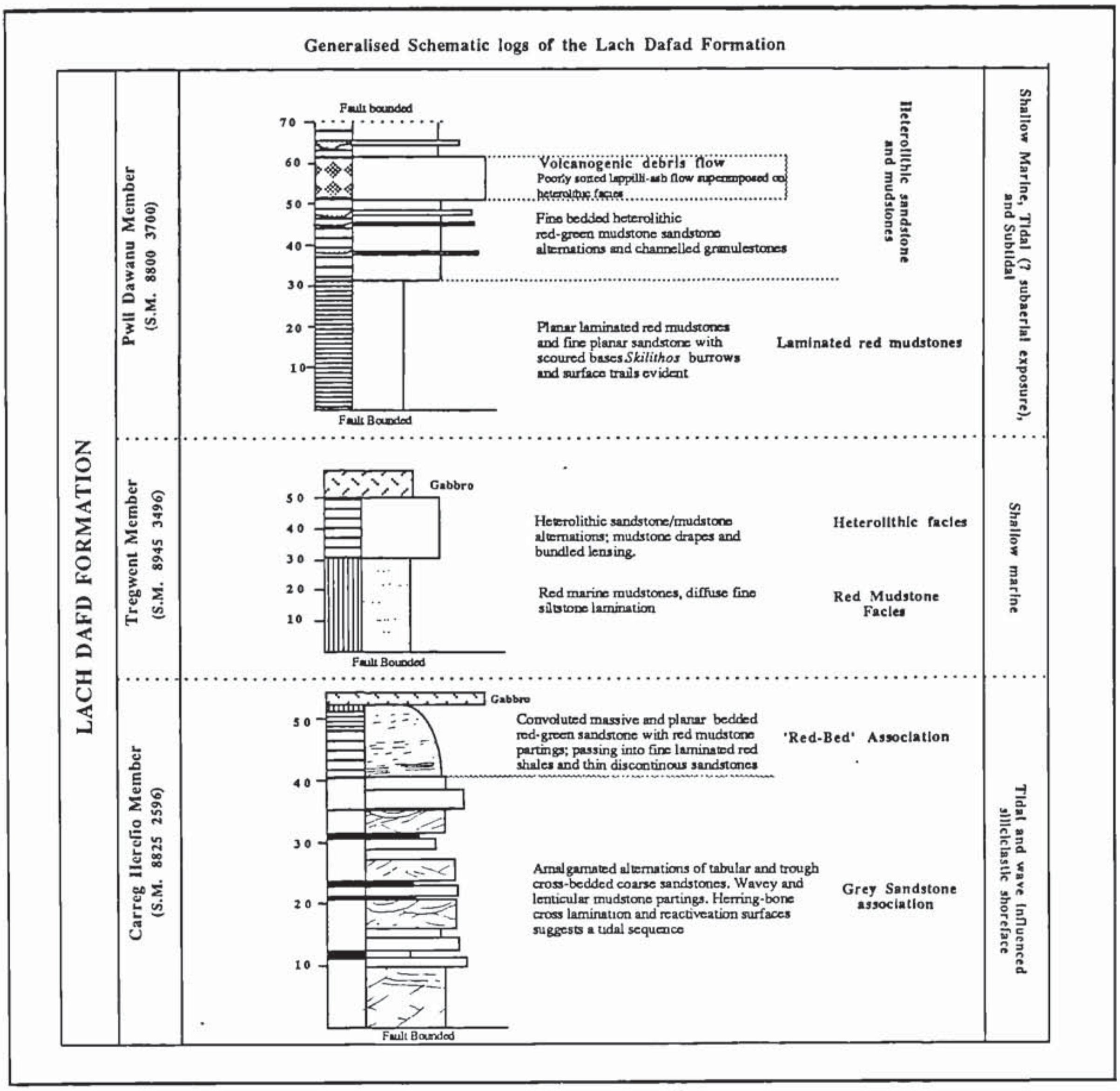

T
V

Table 4.2. Cambrian correlation in SW Dyfed incorporating the fragmented successions of likely Cambrian age developed in the Fishguard – Porth-gain district.

Figure 4.1. Schematic sedimentary log of the Lach Dafad Formation.



CARMARTHEN
ST. DAVIDS
HAYSCASTLE
FISHGUARD PORTH-GAIN
RAMSEY ISLAND
Ordovician
Tremadoc Series
Arenig
Arenig
Arenig
Arenig
Upper Cambrian (Merioneth Series)
Lingula Flags
Disconformity
Dolgelly Beds
Festiniog Beds
Shales
Lingula Flags
Lingula Flags
Disconformity
Lingula Group
Porth Ffynnon Formation
Mynydd Morfa Formtion
Ogof Velvet Formation
Middle Cambrian (St. Davids Series)
Disconformity ?
Menevian Group
Upper Solva Beds
Middle Solva
Purple Sandstone Member
Grey Sandstone Member
Lower Solva Beds
Ford Beds
Grey Sandstone Mb.
Musland Grits
Lach Dafad Formation
Pwlldawnau Member
Tregwent Mb.
Carreg Herefio Member
Carn Llundain Formation
Lower Cambrian (Comley Series?)
Lower Cambrian (Conglomerates)
Caerfai Group
Caerbwdy Member
Caerfai Bay Shale Member
St. Nons Member
Conglomerate
Welsh Hook Beds
Red green Sandstone
Red Shales
Green Sandstones
Conglomerate
Precambrian rhyolites
Pebidian/Dimetian Complex



Generalised Schematic logs of the Lach Dafad Formation
LACH DAFD FORMATION
Pwll Dawanu Member (S.M. 8800 3700)
Fault bounded
70
60
50
40
30
20
10
Volcanogenic debris flow
Poorly sorted lappilli-ash flow superimposed on heterolithic facies
Fine bedded heterolithic red-green mudstone sandstone alternations and channelled granulestones
Heterolithic sandstone and mudstones
Planar laminated red mudstones and fine planar sandstone with scoured bases Skilithos burrows and surface trails evident
Laminated red mudstones
Fault Bounded
Shallow Marine, Tidal (? subaerial exposure), and Subtidal
Tregwent Member (S.M. 8945 3496)
Gabbro
50
40
30
20
10
Heterolithic sandstone/mudstone alternations; mudstone drapes and bundled lensing.
Heterolithic facies
Red marine mudstones, diffuse fine siltstone lamination
Red Mudstone Facies
Fault Bounded
Shallow marine
Carreg Herefio Member (S.M. 8825 2596)
Gabbro
50
40
30
20
10
Convoluted massive and planar bedded red-green sandstone with red mudstone partings; passing into fine laminated red shales and thin discontinous sandstones
'Red-Bed' Association
Amalgamated alternations of tabular and trough cross-bedded coarse sandstones. Wavey and lenticular mudstone partings. Herring-bone cross lamination and reactiveation surfaces suggests a tidal sequence
Grey Sandstone association
Fault Bounded
Tidal and wave influenced siliciclastic shoreface

Herefio Member to strongly suggest deposition within a tidally influenced environment. The Tregwent Member is thought to be of shallow marine affinity, whilst the Pwlldawnau Member is dominated by a complex amalgamation of lithologies not readily discriminated.

There is little doubt in the author's mind and apparently that of Cox (1930), that the Lach Dafad Formation represent the development of either the Lower or Middle Cambrian in this part of SW Dyfed. The present author favours the Middle Cambrian (i.e. Solva Group), though irrespective of specific age, the observations presented complicate the rather vague understanding of the Lower and Middle Cambrian palaeogeography of the area; although little is known of the principle sequences of the Caerfai, Solva and Menevian Groups {2.2.}. At present the general view, is one of deposition along the southern margin of the Welsh Basin (Crimes 1970, Cowie & Rushton 1974), the presence of tidal deposit (Carreg Herefio Member) within the Fishguard – Porth-gain district suggest however, that a marine shoreface was at one time significantly further north than in the classical 'Welsh Basin' model.

If the Lach Dafad Formation is of Middle Cambrian age, successions of the stratotype Lower (Caerfai Group) and Upper Middle Cambrian (Menevian Group) would appear to be absent. Omission by faulting may be the likely cause, although a similar absence is noted on Ramsey Island (Kokelaar *et al.* 1985) whilst around the Carmarthen district, Cope (1979) documents an Upper Middle Cambrian non-sequence which he attributes to uplift and basement faulting (Table 4.2.). Speculatively such uplift may be appropriate to the apparently limited and variable thickness distribution of Lower and Middle Cambrian within SW Dyfed. It being possible that throughout early Cambrian times sediments accumulated in small, although interrelated basins, with structures such as the St. David's and Brawdy-Hayscastle antiforms, with their 'horst-like' geometries and pre-Caledonide structures (Thomas & Jones 1912, Baker 1982), actively controlling the site and rate of sedimentation. There is certainly evidence throughout the Cambrian and early Ordovician sequences to support the view that Precambrian basement rocks formed a consistent source area. This indicates that Precambrian terrains were periodically emergent within the shallow Cambrian seas as a result of eustatic events and/or local basement faulting.

4.3. UPPER CAMBRIAN STRATIGRAPHY AND SEDIMENTATION

4.3.1. *Lingula* Group (LG) *Thickness (> 250m).*

Upper Cambrian sediments are recorded in the Fishguard – Porth-gain district by successions referred to collectively as the *Lingula* Group (the *Lingula* Flags of previous workers). The lithological nature of the Group is seen to vary, although all appear to be of shallow marine affinity, characterised by alternating sequences of fine to medium bedded mudstones, siltstones, and fine to medium sandstones; only in rare circumstances are coarser grained lithologies observed.

Subdivision into the Porth Ffynnon and Mynydd Morfa Formation, distinguishes successions cropping out between Porth-gain and Trwyn Llwyd (Map 4.), and Pwllstrodur and Pwlchrochan (Map 3.). This division is primarily aimed at highlighting the stratigraphic position of the Porth Ffynnon Formation at the top the Cambrian in this part of SW Dyfed. The recorded occurrence of *L. davisii* (Cox 1916, 1930) within sediments of both the Mynydd Morfa and Porth Ffynnon Formations indicates an Upper Cambrian age (see below).

(a). Mynydd Morfa Formation (B1). *Type Locality [S.M. 8730 3480], Thickness (> 250m)*

The Mynydd Morfa Formation is comprised of heterolithic shallow marine clastics, and occupies the northern most part of the Mynydd Morfa headland (Map 3.). The Formation is in excess of 250m thick at this point and represents the thickest development of the Upper Cambrian in the Fishguard – Porth-gain district. It is fault bounded to the south against red-green sandstones of the Lach Dafad Formation and spatially overlain to the north by Arenig sediments of the Trwyn Llwyd Formation; the contact with which is hidden by beach cover in the vicinity of Aber Mawr [S.M. 8830 3460]. The Formation is complexly folded, although continuous exposure in the cliffs and wave-cut platform at Porth Mawr [S.M. 8710 3475] and Penmorfa [S.M. 8730 3480] allows the succession to be studied.

Description: The typical lithology of the Mynydd Morfa Formation is one of complex mudstone-siltstone-sandstone alternations. A complete array of interbedded lithologies and bed forms are developed, varying from: mudstones, streaked mudrocks, lenticular/flaser alternations, and sheet sandstones, all of which are amalgamated on a meter scale, with little, or no, evidence for systematic order (Fig. 4.2.). The amalgamation of lithologies, is thought to represent fluctuating styles of sedimentation within the same depositional environment.

Mudstones with limited sedimentary structure are common, forming individual units that vary in thickness from 5-20cm. Fine silt grade laminae are locally evident, which is frequently discontinuous and convoluted as a result of bioturbation. No systematic variation in bioturbation is evident.

Streaked mudrocks are a coarser grained lithology where siltstone and more rarely fine sandstones form thin discontinuous lenses with irregular swell and pinch bedding surfaces commonly draped by mudstone (Plate 4.5.). Individual beds are variable from the millimetre scale to several centimetres. *Lenticular/Flaser alternations* reflect a coarser variant on the streaked mudrocks, where the sandstone:mudstone ratio is higher. Beds characteristically show wavy alternations where isolated sandstone occur as both lens and laterally continuous bed forms (*Linsen structures*), through various permutations of lenticular and flaser bedded units.

Flaser beds locally approach a point where mud drapes are but a whisp and approximate sheet sandstones. Gutter casts whilst not common, are present within this coarser grained lithology.

Fine-medium grained sheet sandstones occur as discrete planar bedded units interbedded with all the above lithologies (Plate 4.5.). They are characterised by their lateral continuity and textural maturity, and thickness from several centimetres to tens of centimetres. The base of individual beds is commonly planar to weak undulatory, whilst upper surfaces are generally sharp. Internally beds occasionally, exhibit planar lamination although low angle cross lamination is more common.

Interpretation; The amalgamated character of the Mynydd Morfa Formation reflects a shallow environment characterised by rapid fluctuations in energy conditions. The above rock types are not uncommon to a variety of shallow marine environments, although, as outlined by Swift *et al.* (1987), there is as yet no generally accepted convention for equating sedimentary facies as defined by the primary structures and lithology of a rock volume with shallow marine depositional environments (i.e. shoreface, shelf) as defined by the depth and slope of the depositional environment. The simplest explanation of the alternations is that deposition reflected a continuum of processes from high to low energy within a wave dominated inner marine shelf (de Raff *et al.* 1977). Mudstones reflect prolonged fair-weather conditions progressively passing through coarser lithologies where deposition resulted from fluctuating wave intensity, to a point were sheet sandstones reflect either intense wave and/or storm sedimentation.

(b). Porth Ffynnon Formation (B2). *Type locality [S.M. 8114 3265]; Thickness (>130m).*

The Porth Ffynnon Formation occupies the coastal districts between Porth-gain and Trwyn Elen [S.M. 8195 3285] from where successions can be traced onto the headland of Yns-fach [S.M. 8220 3275]. Access is limited at both localities, although the Formation can be clearly seen to be fault bounded against Arenig shales (Penmaen Dewi Formation). North of Trefin, it occupies the headland of Trwyn Llwyd [S.M. 8330 3295] representing the along strike continuation of the Yns-fach exposure (Map 4.); here the Formation is seen to overlie (in complex juxtaposition) arenaceous sediments of the Abercastle Formation (mid-Arenig). To the east of Trwyn Llwyd at Pwl Olfa [S.M. 8365 3295] the top of the Formation is fault bounded by shales of the Penmaen Dewi Formation and uppermost successions of the Abercastle Formation (Map 4). A representative and accessible section which shows the salient lithological characteristics of the Formation, crops out on the headland of Trwyn Porth-gain [S.M. 8114 3265].

The true thickness of the Porth Ffynnon Formation is unclear. An estimated outcrop thickness of 130m is thought appropriate, although the base of the succession is no where seen, and folding complicates much of the exposure. The top of the Formation is observed at Porth-gain Harbour where it is seen to be overlain with significant hiatus by Arenig strata, referred to here as the Porth-gain Harbour Formation (the 'Porth Gain Beds' of Cox 1916 and the 'Porth Gain Formation' of Fortey & Owens 1987, {4.5.}). Transition between the Porth Ffynnon Formation and the Porth-gain Harbour Formation is observed at the east-side of the Porth-gain harbour entrance north of the jetty [S.M. 8140 3265]. Here polymictic breccias highlight the hiatus between the uppermost Cambrian (?Maentwrog Stage) and lowermost Ordovician (Lower Arenig – Moridunian Stage) marine sediments.

Description: Sediments of the Porth Ffynnon Formation differ from the Mynydd Morfa Formation to the west, in as much as they are dominated by alternations of *siltstones* and thin bedded *texturally mature 'sheet' sandstones* with little admixture of grain size between the two variants (compare Plates 4.5. & 4.6.). However, both lithologies can be regarded as the most frequent rock type within a complex depositional sequence.

Siltstones may show faint parallel and small wavy ripple lamination. Thin discontinuous ormally graded coarse sandstone and granulestone laminations are locally evident. From the sections studied, no direct evidence for bioturbation was found. However, this is not to infer that the Formation is barren of biogenic influence, the homogeneous nature and 'sugary' t xture to much of the silt suggests complete destruction of sedimentary structures through b oturbation. Discrete mudstones are rare.

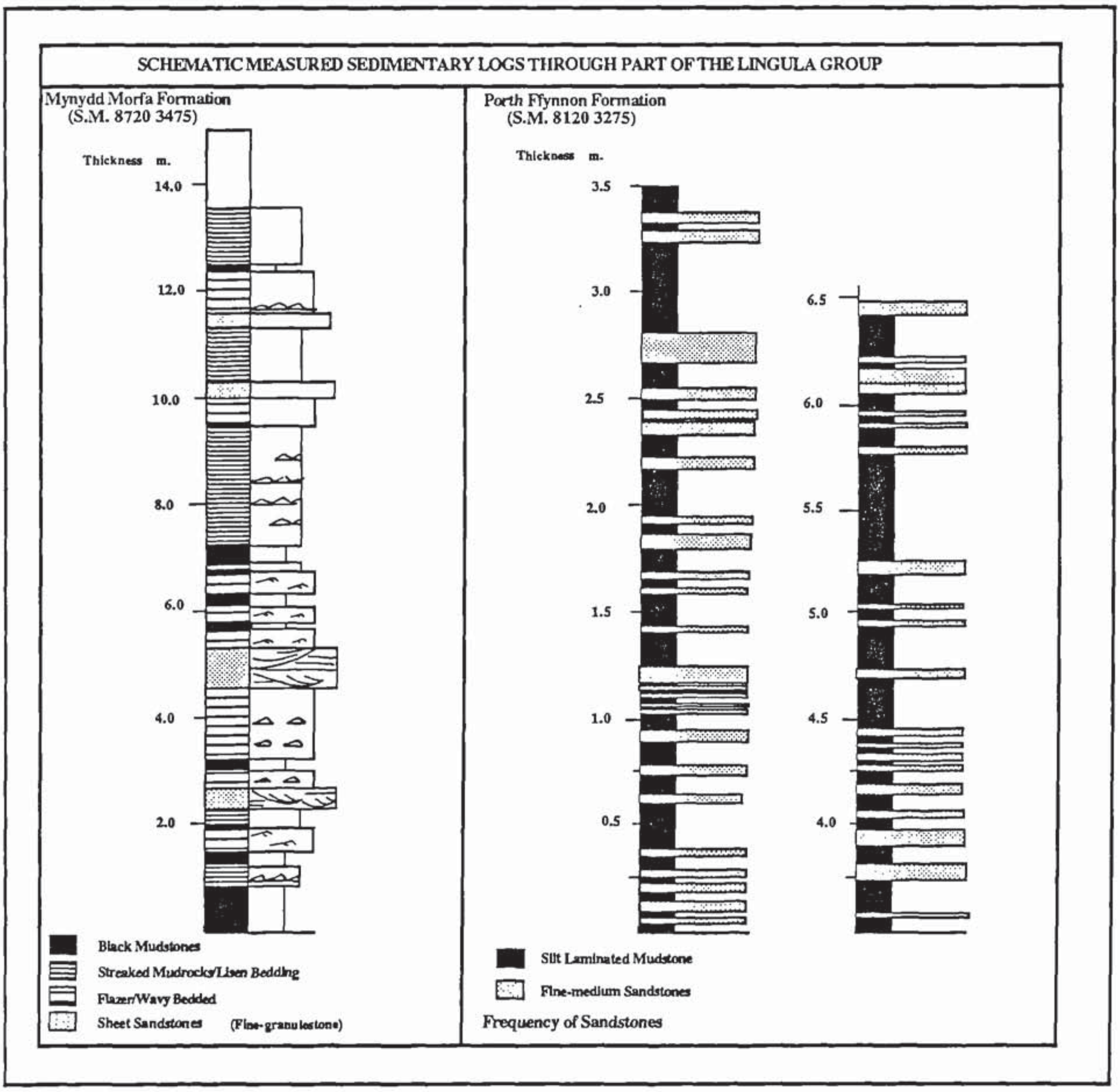
The *texturally mature 'sheet' sandstones* vary in thickness from several centimetres to ≈20cm; rarely are beds thicker than this seen. The sandstones are characterised by good lateral continuity through outcrop (>100m), although locally discontinuity is common as a result of pinch and swell bedding surfaces. The sandstones are well sorted and fine to medium grained. Externally the geometry of individual sandstone beds is varied. Lower bounding surfaces are generally sharp, although vary from planar to undulatory (Plate 4.6.); the latter reflects both the pre-existing morphology and scoured/erosive contacts. Upper surfaces vary from planar to weak undulatory, through to symmetric and asymmetric rippled geometries. Internally sandstones show a wide range of sedimentary structures, comparable to those recorded by Brenchley and Newall (1981) from the Cherney Llongville Flags, Shropshire. Parallel lamination is common at the bottom of individual planar beds and may pass upwards into cross-laminated sandstones. Low-angle cross lamination and hummocky cross-stratification is common in thicker beds (Plate 4.6.). All of these are features typical of sublittoral sheet sandstone (Goldring & Bridges 1973).

Plate 4.5. Typical alternation of lithology within the Mynydd Morfa Formation varying from streaked mudrocks through to sheet sandstones (S). The dark lithology at the top of the plate is a dolerite sill. [S.M. 8730 3480]

Plate 4.6. Texturally mature 'sheet' sandstones and interbedded siltstones of the Porth Ffynnon Formation. Note that sandstone geometry is dictated in part by the pre-existing morphology. The sandstone centre view shows low angle hummocky cross-lamination


S

Fig. 4.2. Schematic sedimentary logs of the Porth Ffynnon and Mynydd Morfa Formations showing the variability in lithologies and frequency of sandstones.



SCHEMATIC MEASURED SEDIMENTARY LOGS THROUGH PART OF THE LINGULA GROUP
Mynydd Morfa Formation
(S.M. 8720 3475)
Thickness m.
14.0
12.0
10.0
8.0
6.0
4.0
2.0
Black Mudstones
Streaked Mudrocks/Lisen Bedding
Flazer/Wavy Bedded
Sheet Sandstones (Fine-granulestone)
Porth Ffynnon Formation
(S.M. 8120 3275)
Thickness m.
3.5
3.0
2.5
2.0
1.5
1.0
0.5
6.5
6.0
5.5
5.0
4.5
4.0
Silt Laminated Mudstone
Fine-medium Sandstones
Frequency of Sandstones

Interpretation: The alternations of fine- to medium sheet sandstones within fine laminated siltstones is most easily explained in terms of 'inner shelf' sedimentation. Where the repetitive juxtaposition of lithologies reflects contrasting energy states during 'storm' and 'fair-weather conditions (Mount 1981). The complex internal structures of sandstone beds can be attributed to wave and storm related processes, extensively documented in the literature (see Swift et al. 1983, Duke 1984). For example, low angle cross-lamination is likely to reflect oscillatory wave motions, whilst features such as scoured bases and planar lamination are most easily explained in terms of plane-bed deposition from currents. Collectively these are thought to reflect discrete high energy events during prolonged periods of silt deposition. The source of the sand may be from the shoreline or near shore environment, transported off shore by traction currents during storm surges (Brenchley & Newall 1981), with subsequent wave modification during the waning stages of storm activity. It can be visualised that sedimentation within the Porth Ffynnon Formation was taking place on a muddy marine shelf which was influenced by both wave and/or storm related processes.

Relationships between the Mynydd Morfa and Porth Ffynnon Formation

The Porth Ffynnon Formation is overlain with hiatus by sediments of the Porth-gain Harbour Formation, which are regarded as of lowermost Arenig age {4.4.}. This would suggest (allowing for indeterminate erosion) that the Porth Ffynnon Formation occupies a position at the top of the Upper Cambrian as developed in this part of SW Dyfed (Table 4.2.). It seems likely that the Mynydd Morfa Formation occupies a slightly lower stratigraphic position, although given that the Formation is in close proximity to Arenig strata at Aber Mawr (Map 3.), its uppermost beds may well be equated with the Porth Ffynnon Formation.

Both the Mynydd Morfa and Porth Ffynnon Formations are storm and wave influenced sequences and whilst not being identical lithologically, depositional processes were characterised by rapid energy fluctuations. It seems likely that throughout the Upper Cambrian in this part of SW Dyfed, deposition took place over a stable siliciclastic shelf.

Age constraints and regional correlation of the Lingula Group

The duration of the non-sequence between the Cambrian and Ordovician as recognised by the sedimentary hiatus between the Porth Ffynnon Formation and Porth-gain Harbour Formation {4.4.} is difficult to assess. It is recognised that *L. davisii* is typical of the North Wales Festiniog beds (i.e. mid-Upper Cambrian), although Lake (1906, 1908) and Cowie *et al.* (1972) record *L. davisii* in association with *Homagnostus obesus* (Belt), *Olenus cataractes* Salter, and *O. mundus* Lake (from Hayscastle and Trefgarne), faunas typical of the *Olenus & Homagnostus* Biozone of the Maentwrog Stage (i.e. lower Upper Cambrian). There is no faunal evidence for beds of Festiniog or Dolgelly age in SW Dyfed (Rushton 1974). Whilst this does not necessarily preclude their presence, if Dolgelly and possibly Festiniog Stage sediments are absent, then the non-sequence between the Cambrian and Ordovician is likely to have been of significant duration. SW Dyfed may therefore have formed a positive topographic area during the Festiniog/Dolgelly Stages of the Upper Cambrian and throughout the Tremadoc. Subsequent Ordovician deposition resulted during early Arenig times in response to the basal Arenig transgression. The inference of protracted uplift has implications for the nature of the basinal setting, as it has been recently demonstrated that Festiniog-Dolgelly and Tremadoc sediments were being deposited immediately to the east around Carmarthen (Cope 1982, Fortey & Owens 1987). It seems likely that both areas were distinct terrains influenced by both localised tectonics, and larger 'basin-wide' events.

The *Lingula* Group (the *Lingula* Flags of other workers) is however, poorly understood biostratigraphically, and regional development and depositional histories are for the most part undocumented. However, recent studies of the *Lingula* Group south of St. Davids (Turner 1977) and on Ramsey Island (Kokelaar *et al.* 1985; their Ogof Velvet Formation), may suggest that the majority of the Group was deposited within a relatively shallow wave and current agitated marine environment; although deeper marine conditions may be appropriate above the Menevian Group (Williams & Stead 1982). The local thickness (>600m; Williams 1934) and widespread development from the Prescelly Hills (Lowman 1977) east to Carmarthen (Cope 1979) and south to Ramsey Island (Table 4.2.) suggests that during early Upper Cambrian times SW Dyfed was the site of a large and relatively stable siliciclastic shelf covered by a shallow Cambrian sea. This is in contrast to North Wales where thick deeper marine turbidite deposition took place (Cowie & Rushton 1974).

It is of interest to note that the widespread development of the *Lingula* Group in terms of salient lithologies and aerial, extent is in character with ancient siliciclastic shelves. With regard to this, it is commonly assumed that the southern margin of the Welsh Basin is represented by the Towy Lineament in the east and the present Lower Palaeozoic cover to the west. Spatial reasoning also suggests that area to the south (Bristol Channel) was likely to have existed as a prolonged topographic high. However, from the recent documentation of thick Lower Palaeozoic successions underlying the Bristol Channel (see Tunbridge 1986 *and references therein*), there appears to be little reason why the 'Welsh Basin', during Upper Cambrian times, did not extend significantly further to the south. The successions of South Dyfed therefore representing part of a far larger 'marginal' terrain.

4.4. LOWER – MIDDLE ARENIG STRATIGRAPHY AND SEDIMENTATION

The Arenig of the Fishguard – Porth-gain districts is represented by thick shallow marine clastics which pass upwards into marine mudstones (i.e. Arenig-Llanvirn Shales; Table 4.3.b.) which span the late Arenig and Llanvirn. The following section describes the lower clastic sequence which is believed to represent the development of both the Moridunian and Whitlandian Stages. Three principle stratigraphic divisions are developed, namely; the Porth-gain Harbour, Abercastle, and Trwyn Llwyd Formations.

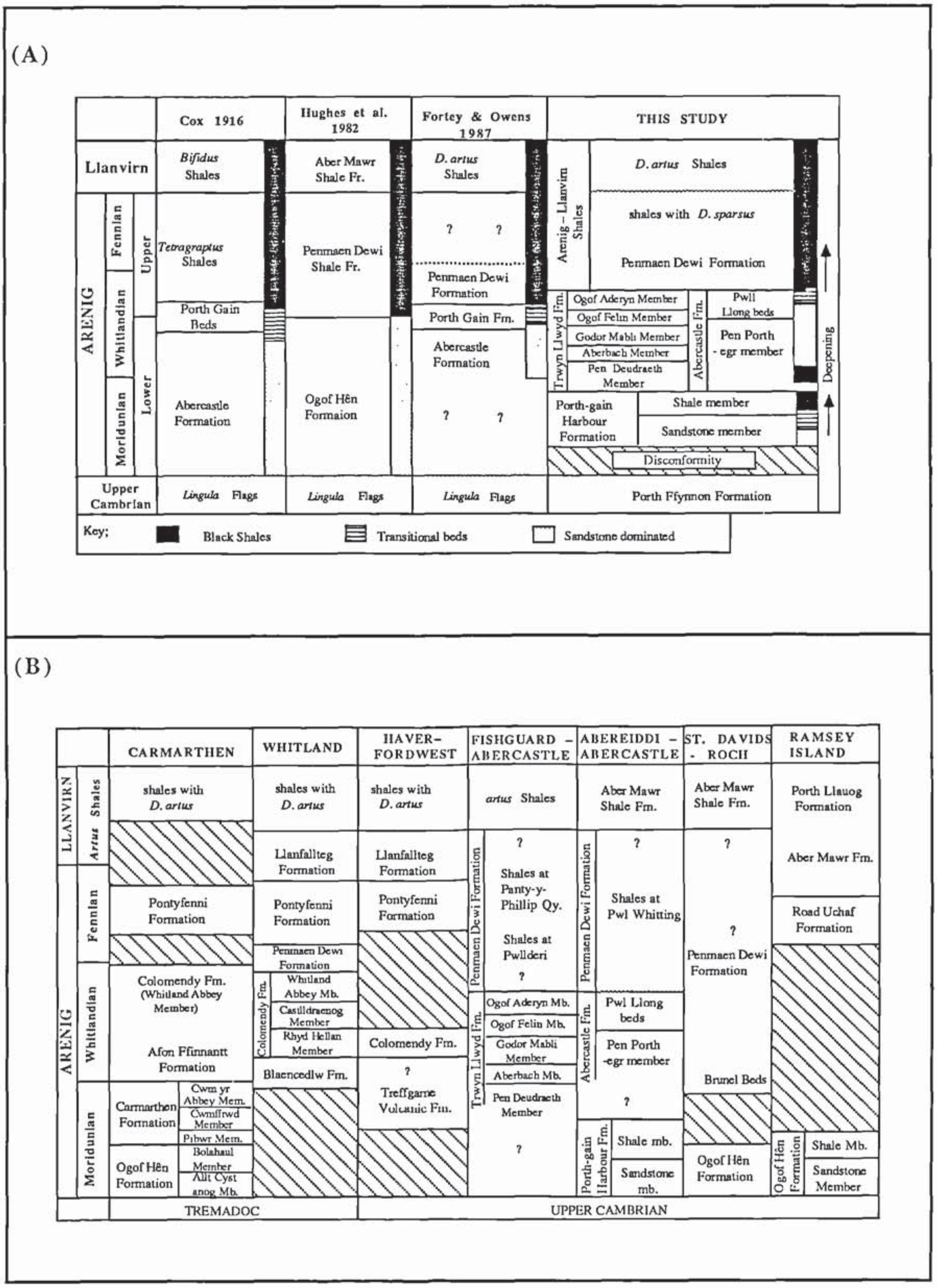
Previous and present interpretations of the Arenig stratigraphy: The Arenig of the Fishguard–Abercastle-Abereiddi district was first detailed by Cox (1916, 1930) who defined the sequence in terms of a lower arenaceous and upper argillaceous division, the Abercastle (Lower Arenig) and *Tetragraptus* Shales (Upper Arenig) respectively; the transition between which was represented by his Porth-gain beds, a member at the top of the Abercastle Formation. Fortey & Owens (1987) whilst agreeing with such lithologic division, place the entire sequence on faunal grounds within the Whitlandian Stage (Fig. 4.3.a.). Hughes *et. al.* (1982) comment on the stratigraphy of the Arenig north of Abereiddi Bay, but do not consider the sequence developed further to the west.

In this study the lower 'arenaceous' Arenig is regarded as being more complex than simply a transition from clastic to argillic sedimentation. The Porth-gain Harbour Formation as defined herein, whilst equating in part with Porth-gain beds of previous workers, is thought by virtue of its 'discordant' relationship upon the Upper Cambrian, to reflect the onset of Ordovician sedimentation in this part of SW Dyfed. The Formation's lithological character shows similarities with the Ogof Hên Formation developed elsewhere in SW Dyfed {2.2.}; on the basis of which it is realistically placed in the Moridunian.

The term Abercastle Formation is retained, although the term Porth-gain beds (or Formation) is abandoned. The Formation is subdivided informally into the Pen Porth-egr member and Pwll Llong beds. The subdivision is aimed primarily at describing a lower clastic sequence and an upper laminated argillic sequence, the latter reflecting the transition into the overlying Penmaen Dewi Formation. The Abercastle Formation is proven to be of Whitlandian age (Fortey & Owens 1987).

The term Trwyn Llwyd Formation is newly introduced to describe a thick siliciclastic-argillic shelf sequence developed between Pwll Deri and Pwllstrodur (Map 3). The Formation is nowhere exposed in it's entirety, although lithological division into 5 members is presented. Based on various considerations the Formation is argued to be of Whitlandian age {4.4.3.}, and, with the Abercastle Formation may represent part of a complex and possibly extensive mid-Arenig shallow marine sequence, which formed part of northward propagating clastic wedge, prior to 'stagnation' and deposition of the Arenig-Llanvirn Shales {4.5.}.

Table 4.3. **(A)** Previous and present interpretations of the Arenig successions in SW Dyfed (St. Davids – Fishguard). **(B)** Correlation of Arenig strata in SW Dyfed.


(A)
Cox 1916
Hughes et al. 1982
Fortey & Owens 1987
THIS STUDY
Llanvirn
ARENIG
Fennian
Whitlandian
Moridunian
Upper
Lower
Upper Cambrian
Bifidus Shales
Tetragraptus Shales
Porth Gain Beds
Abercastle Formation
Lingula Flags
Aber Mawr Shale Fr.
Penmaen Dewi Shale Fr.
Ogof Hên Formaion
D. artus Shales
?
Penmaen Dewi Formation
Porth Gain Fm.
Abercastle Formation
Arenig – Llanvirn Shales
D. artus Shales
shales with D. sparsus
Penmaen Dewi Formation
Trwyn Llwyd Fm.
Ogof Aderyn Member
Ogof Felin Member
Godor Mabli Member
Aberbach Member
Pen Deudraeth Member
Abercastle Fm.
Pwll Llong beds
Pen Porth - egr member
Porth-gain Harbour Formation
Shale member
Sandstone member
Disconformity
Porth Ffynnon Formation
Deepening
Key;
Black Shales
Transitional beds
Sandstone dominated
(B)
CARMARTHEN
WHITLAND
HAVER-FORDWEST
FISHGUARD – ABERCASTLE
ABEREIDDI – ABERCASTLE
ST. DAVIDS - ROCH
RAMSEY ISLAND
LLANVIRN
Artus Shales
ARENIG
Fennian
Whitlandian
Moridunian
shales with D. artus
artus Shales
Aber Mawr Shale Fm.
Porth Llauog Formation
Llanfallteg Formation
Aber Mawr Fm.
Pontyfenni Formation
Road Uchaf Formation
Penmaen Dewi Formation
Shales at Panty-y-Phillip Qy.
Shales at Pwllderi
Shales at Pwl Whitting
Colomendy Fm. (Whitland Abbey Member)
Afon Ffinnantt Formation
Colomendy Fm.
Whitland Abbey Mb.
Castilldraenog Member
Rhyd Hellan Member
Blaencedlw Fm.
Treffgarne Volcanic Fm.
Trwyn Llwyd Fm.
Ogof Aderyn Mb.
Ogof Felin Mb.
Godor Mabli Member
Aberbach Mb.
Pen Deudraeth Member
Abercastle Fm.
Pwl Llong beds
Pen Porth -egr member
Brunel Beds
Carmarthen Formation
Cwm yr Abbey Mem.
Cwmffrwd Member
Pibwr Mem.
Ogof Hên Formation
Bolahaul Member
Allt Cyst anog Mb.
Porth-gain Harbour Fm.
Shale mb.
Sandstone mb.
Ogof Hên Formation
Shale Mb.
Sandstone Member
TREMADOC
UPPER CAMBRIAN
?

4.4.1. The Porth-gain Harbour Formation (C). *Type Locality [S.M. 8140 3265]; Thickness (≈45m)*

The Porth-gain Harbour Formation represents a thin succession of heterolithic shallow marine sediments, which overlie with hiatus, sediments of the Porth Ffynnon Formation (*Lingula* Group) at the east side of the entrance to Porth-gain Harbour [S.M. 8140 3265, Map 4.]. This locality relates directly to the stratotype locality of the Porth-gain beds of previous workers. The Formation is observed only at this locality, where it is seen to fine-up from conglomerates and sandstones into mudstones. It is convenient for stratigraphic purposes to discuss the Formation informally in terms of this lithologic division (**Sandstone member** and **Shale member**, Fig. 4.4.), although the transition between both members is gradational, probably reflecting increasing water depth.

a) Sandstone member (≈ 15-20m)

The Sandstone member represents a thin succession of polymictic conglomerates and breccias, sandstones, and interbedded mudstones. The base of the member is observed in the shallow cliffs approximately 40m to the north of the jetty on the east side of the entrance to Porth-gain Harbour [S.M. 8140 3625], where it is seen to overlie the Porth Ffynnon Formation with hiatus. The exact transition between both Formations can not be 'pinpointed' to a specific bed, rather to a sequence of thin weakly channelled conglomerates/breccias and diffuse tabular crossbedded sandstones. The relationship between both Formations shows no evidence for an angular discordance, although the entire cliff-section shows evidence for shearing and limited bedding planar movements. Nevertheless, the present author contends that this locality, reflects the transition between Cambrian and Ordovician marine strata in this part of SW Dyfed.

Description: Sediments of the Sandstone member can be best considered in two broad groups (Fig. 4.4.); a lower sequence of immature coarse sandstones and breccias, which pass upwards into heterolithic mudstone/sandstone alternations. The latter pass gradationally into mudstones of the Mudstone member (see below).

Coarse sandstones and sedimentary breccias which form the base of the Sandstone member are compositionally and texturally immature and poorly sorted. Sandstones dominate the sequence within which stratification is for the most part difficult to resolve. Where mudstone and graded lamination is sufficiently highlighted by preferential weathering, it is possible to define bedding which shows some evidence for weak tabular cross-sets locally highlighted by thin discontinuous muddy 'flazers', suggesting reactivation surfaces. Bioturbation is not observed. Within the sandstones, angular to sub-rounded cobbles and boulders with no apparent order or imbrication sit isolated within the poorly resolved matrix. The nature of clasts is seen to vary, although the dominant type are weakly vesicular rhyolitic ash-flow tuff and jasperiod volcanics.

At discrete horizons, poorly sorted, polymictic breccias occur interbedded within the sandstones. Individual beds rarely seen to exceed 20cm in thickness and are generally discontinuous. They commonly possess an asymmetrically scoured channel profile; top surfaces are planar, although locally show evidence for tectonic modification. The breccias are framework supported showing a wide grain size distribution from medium- to coarse sandstone through to pebble grade. Weak normal grading is evident, and symmetric grading (normal to reverse) is not uncommon, possibly reflecting 'fines' depletion by winnowing of the top surfaces. Rare cross-bedding is evident, although widespread imbrication was not seen. Clast types are highly variable, and include porphyritic acidic volcanics, quartz-feldspathic porphyries, vein quartz, jasperiod volcanics, intermediate volcanics, weakly strained pssamitic clasts, intra-formational mudstones and lithic arenites similar in character to the underlying *Lingula* Group.

Heterolithic mudstone/sandstone alternations dominate the middle parts of the Porth-gain Harbour Formation (Fig. 4.3.). The general character of the lithology is one of discrete intercalated units varying in thickness from 5-30cm, within which a variety of sedimentary structures are evident. The most common features of the sequence are lenticular, wavy, flazer, and sheet sandstone alternations reflecting rapid fluctuations in the depositional style.

Sheet sandstones may be viewed as the extreme end-member, where the environment was sufficiently vigorous so as not to allow the preservation of muds. This is supported by the presence of small tabular and wedge cross-sets and an overall coarser grain size which varies from fine-medium sandstone to granulestone. Flaser bedded units can be viewed as a slightly lower energy state, were mudstone from 'slack water', whilst not accumulating in any great quantity was locally preserved as isolated drapes and bifurcating wavy lamination; the latter commonly mantling subjacent ripples (Plate 4.7.). Wavy and lenticular bedded units reflect the lower energy state where mudstones and sandstone accumulated over prolonged periods. Bioturbation is evident within these muddier horizons, reflecting less 'hostile' conditions. Channelled orthoconglomerates occur sparodically within the lenticular and wavy bedded alternations, varying in thickness from 10-25cm. Clast types are polymictic, although are similar in character to the underlying breccias. However, clast shape is sub-rounded to rounded, suggesting prolonged retention in the marine environment.

Interpretation: The Sandstone Member is interpreted as a transgressive marine sequence which on-lapped the Upper Cambrian. The diffuse nature of the lower *coarse sandstones and breccias* precludes a detailed interpretation, although there is little doubt from the character of the overlying heterolithic alternations that the sequence is shallow marine; the large scale tabular cross-beds may form part of a larger channelled sequence. The immature nature of the lithology and angular nature of clast types suggests relatively rapid deposition and limited retention in the sedimentary cycle or proximity to source area, or both. The exact origin of the clasts remain uncertain although they are likely to be derived in part from a Precambrian terrain, possibly resulting through transgressive reworking of underlying Precambrian basement rocks; although the position of such a terrain is uncertain. Bevins (*pers comm. in* Fortey & Owens 1987) suggests that the conglomerates are likely to be derived from the reworking of the andesitic Trefgarne Volcanic Group, situated 25km to the southeast from the Porth-gain district. Whilst acknowledging that intermediate volcanic rock fragments are present, the nature and angularity of the clast types suggests a more 'proximal' source is the case.

The *heterolithic* facies is one of composite lithologies reflecting variable hydrodynamic conditions and rapid fluctuations. The alternating wavy, lenticular and flazer bedded units, reflect an intertidal environment within which sediment was deposited from rapidly fluctuating turbulent and slack water conditions. Coarse rippled sandstones may represent single depositional units, deposited during intense storm weather conditions with mud placed into suspension forming muddy flasers. Such flazers could be either reworked during subsequent events forming homogenous sheet sandstones, alternatively, if interstorm periods where of sufficient duration, left as bifurcating drapes (Plate 4.7.). The conglomerates may reflect small outwash deposits, although unlike the underlying breccias they w ere retained sufficiently long in the shallow marine environment to inherit a high degree of sphericity.

b). Shale member (> 25m).

The heterolithic facies of the Sandstone member, gradationally give way to homogenous mudstones referred to here as the Shale member (Fig. 4.6.). The top of the member is not seen, although is likely to be fault bounded against sediments of the Abercastle Formation or shales of the Penmaen Dewi Formation which crop out immediately to the south of Porth-gain Harbour (Map 4.). The thickness of the Shale member as observed at outcrop on the east site of the entrance to Porth-gain harbour is greater than 25m. It is likely that the marine shales cropping out in the harbour floor represent part of the same succession, although they may represent 'faulted-in' younger shale sequences of uppermost Arenig or Llanvirn age. Lithologically the member is generally composed of monotonous mudstones, fine cross- and parallel laminated siltstones are locally evident, although the depositional environment is not readily discriminated by physical characteristics. There is little doubt however, that they reflect a continuum of sedimentation from the underlying heterolithic alternations, possibly reflecting a slight deepening of the marine environment, although no great depth below effective wave base is implied.

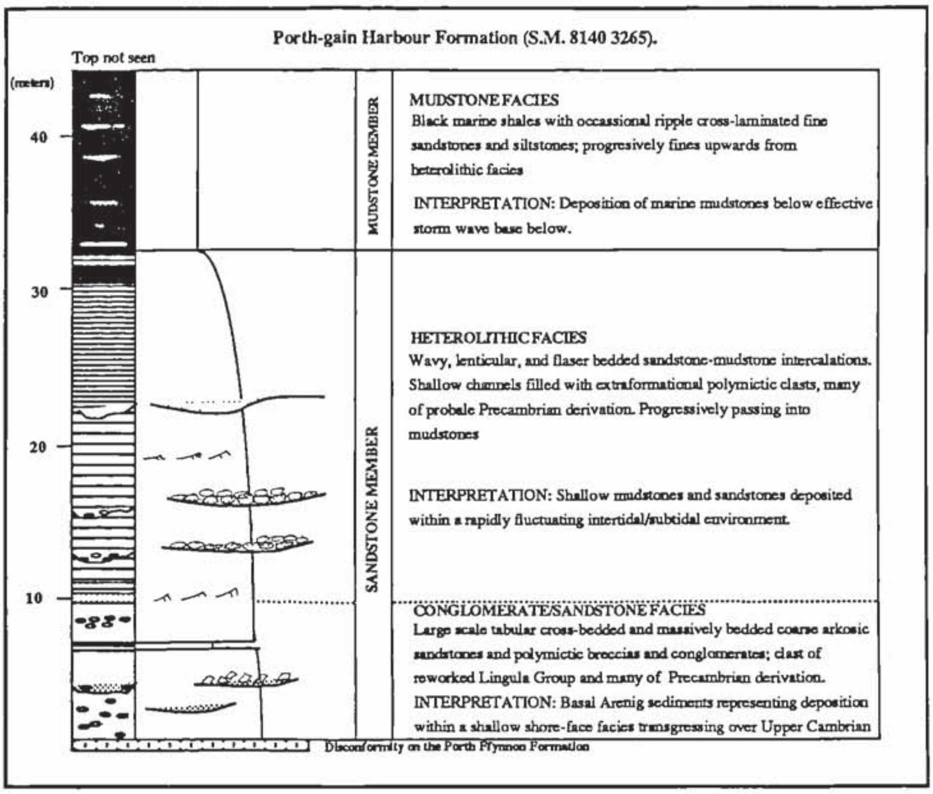
Plate 4.7. Sheet sandstones and wavy, flazer, lenticular bedded alternations within the Sandstone member of the Porth-gain Harbour Formation. These commonly associated bedded forms reflect rapidly fluctuating conditions within a shallow marine intertidal – sub-tidal environments. Note minor bioturabtion at top of plate. Scale coin (arrowed is 2cm in diameter)

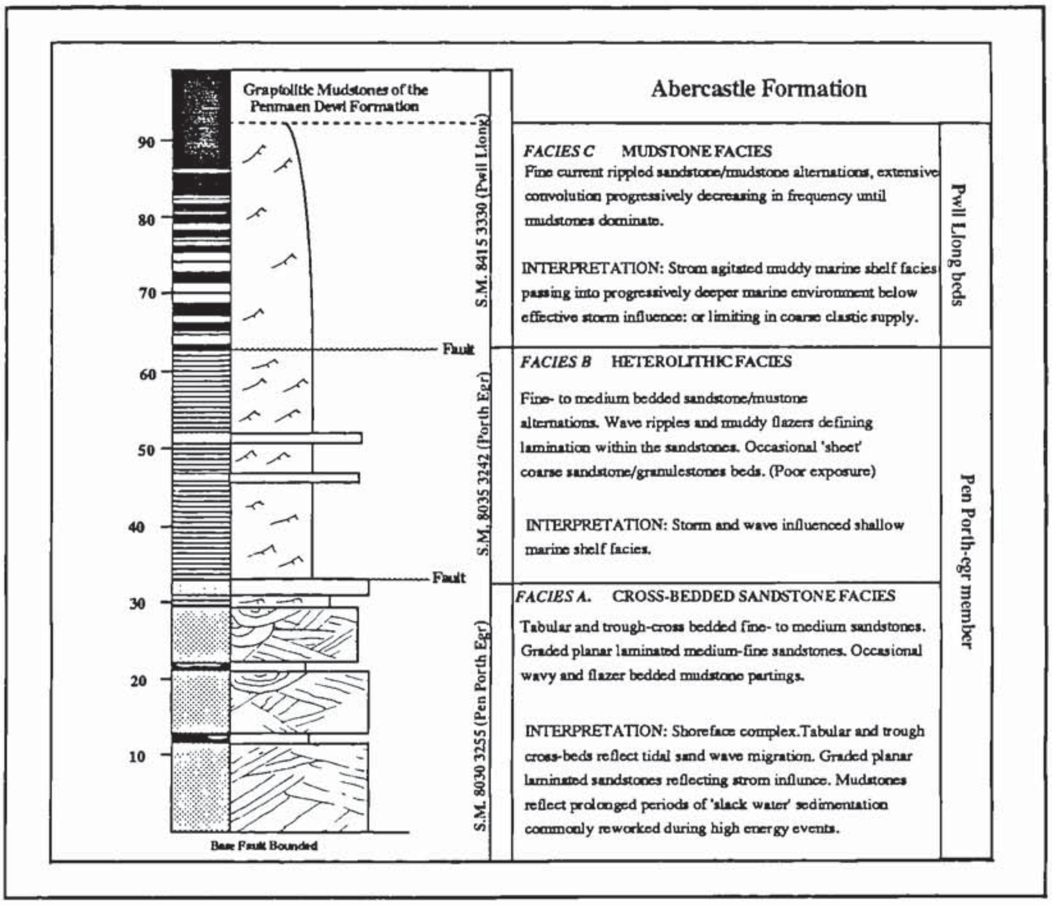
Plate 4.8. Well sorted fine grained tabular and planar laminated sandstones of the Abercastle Formation (Pen Porth-egr member), interpreted as shoreface deposits [S.M. 8030 3250]. Note high and low angle discordances and opposing sense of dips in the tabular cross-sets, and grading of planar lamination (P).


P
CHINON

Figure 4.3. Schematic sedimentary log of the Porth-gain Harbour Formation

Figure 4.4. Schematic sedimentary log of the Abercastle Formation.


Porth-gain Harbour Formation (S.M. 8140 3265).
Top not seen
(meters)
40
30
20
10
MUDSTONE MEMBER
SANDSTONE MEMBER
MUDSTONE FACIES
Black marine shales with occassional ripple cross-laminated fine sandstones and siltstones; progresively fines upwards from heterolithic facies
INTERPRETATION: Deposition of marine mudstones below effective storm wave base below.
HETEROLITHIC FACIES
Wavy, lenticular, and flaser bedded sandstone-mudstone intercalations. Shallow channels filled with extraformational polymictic clasts, many of probale Precambrian derivation. Progressively passing into mudstones
INTERPRETATION: Shallow mudstones and sandstones deposited within a rapidly fluctuating intertidal/subtidal environment.
CONGLOMERATE/SANDSTONE FACIES
Large scale tabular cross-bedded and massively bedded coarse arkosic sandstones and polymictic breccias and conglomerates; clast of reworked Lingula Group and many of Precambrian derivation.
INTERPRETATION: Basal Arenig sediments representing deposition within a shallow shore-face facies transgressing over Upper Cambrian
Disconformity on the Porth Ffynnon Formation


Abercastle Formation
Graptolitic Mudstones of the Penmaen Dewi Formation
90
80
70
60
50
40
30
20
10
Fault
Fault
Base Fault Bounded
S.M. 8415 3330 (Pwll Llong)
S.M. 8035 3242 (Porth Egr)
S.M. 8030 3255 (Pen Porth Egr)
FACIES C MUDSTONE FACIES
Fine current rippled sandstone/mudstone alternations, extensive convolution progressively decreasing in frequency until mudstones dominate.
INTERPRETATION: Strom agitated muddy marine shelf facies passing into progressively deeper marine environment below effective storm influence: or limiting in coarse clastic supply.
Pwll Llong beds
FACIES B HETEROLITHIC FACIES
Fine- to medium bedded sandstone/mustone alternations. Wave ripples and muddy flazers defining lamination within the sandstones. Occasional 'sheet' coarse sandstone/granulestones beds. (Poor exposure)
INTERPRETATION: Storm and wave influenced shallow marine shelf facies.
FACIES A. CROSS-BEDDED SANDSTONE FACIES
Tabular and trough-cross bedded fine- to medium sandstones. Graded planar laminated medium-fine sandstones. Occasional wavy and flazer bedded mudstone partings.
INTERPRETATION: Shoreface complex. Tabular and trough cross-beds reflect tidal sand wave migration. Graded planar laminated sandstones reflecting strom influnce. Mudstones reflect prolonged periods of 'slack water' sedimentation commonly reworked during high energy events.
Pen Porth-egr member

Regional Correlation

The Porth-gain Harbour Formation as defined herein, is regarded as lowermost Moridunian age, comparable to sediments of the Ogof Hên Formation developed elsewhere in SW Dyfed (Table 4.3.). Evidence supporting such correlation and the inference of the succession being of lowermost Arenig age rather than mid-Arenig (Table 4.3.), is founded on the following points of argument:

Firstly, sediments of the Porth-gain Harbour Formation rest with significant sedimentary hiatus on the Porth Ffynnon Formation expressed by reworking of the underlying Upper Cambrian and an extraformational source (not too dissimilar from the Precambrian Pebidian–Dimetian Complex of the St. Davids district {2.2.}). They therefore represent the earliest Ordovician strata in this part of SW Dyfed. The Formation reflects deposition within a rapidly transgressing marine sequence, typical of the lowermost Arenig elsewhere in the Welsh Basin (Zalasiewicz 1984, Legget 1982, Beckley 1987; Fortey & Cocks 1986, Kokelaar *et al.* 1985, Fortey & Owens 1987). The overall lithological characteristics of the Formation are akin to the Ogof Hên Formation developed in SW Dyfed (i.e. thin polymictic breccias passing rapidly into mudstones), particularly the successions developed on Ramsey Island (Kokelaar *et al.* 1985). It is of interest to note, that phosphatic nodules have recently been identified in sediments referred to here as the Porth-gain Harbour Formation (*pers. comm.* J.J. Taynor 1987). indicating a period of non-deposition (*pers. comm.* Dr. A. J. Hartley 1988). Phosphatic deposits are common to other basal Arenig successions, particularly on Ramsey Island where Hofmaan (1975) identifies phosphatic oncoliths (previously referred to as the bryozoan *Bolopora undosa*) within the basal beds of the Ogof Hên Formation.

It would appear that the fining-up character (Sandstone member – Shale member) of the Porth-gain Harbour Formation has been the principle argument upon which previous workers have based their stratigraphic subdivision of the Arenig in this area (Table 4.3.; *n.b.* the fault that Cox (1930) describes as further evidence occurs ≈25m to the south of the sedimentary hiatus). However, it is apparent from the Trwyn Llwyd Formation {4.4.3.} that thick black marine shales were being deposited in a shelf environment prior to the development of shallow marine conditions during the Whitlandian (i.e. mid-Arenig).

This points to an area of complexity hitherto unappreciated in so much as it would appear that early Arenig deposition resulted firstly during relatively rapid transgression (i.e. Porth-gain Harbour Formation). Subsequent shallowing and probable shelf progradation followed during mid-late Moridunian times with a return to shallow marine conditions during the Whitlandian (i.e. Abercastle and Trwyn Llwyd Formation). This subsequently gave way to the development of the Penmaen Dewi Formation (i.e two transgressive events, Table 4.3.). Such lithostratigraphic interpretation whilst complex, can be viewed as in character with the Arenig elsewhere, in so much as Fortey & Owens (1987) suggest that early Arenig deposition in the Carmarthen district is recorded by two 'upward-deepening' successions.

4.4.2. Abercastle Formation (D). *Thickness (> 90m).*

The term Abercastle Formation was introduced by Cox (1916) for the arenaceous Lower Arenig successions developed around Abercastle and Porth-gain (Table 4.3.). Based on the recognition of *Ogyginus hybridus* (referred to as *'O. selwynii'* by Cox 1916) the Formation is reassigned to the mid-Arenig Whitlandian Stage (Fortey & Owens 1987). Both Cox (1916) and Fortey & Owens (1987), considered the 'lithotype' Porth-gain Harbour Formation as described herein (their Porth-gain Formation or beds), to represent a member at the top of the Abercastle Formation (Table 43.a.). Whilst their Porth-gain beds are here inferred to be of Moridunian age and represent the basal Arenig as developed in this part of SW Dyfed, it is only at the lithotype locality where interpretations differ; there being little doubt that the Abercastle Formation elsewhere, reflects a steadily deepening marine sequence from arenaceous strata into mudstones of the Penman Dewi Formation (Table 4.3.a.).

The Abercastle Formation is nowhere exposed in its entirety, although a thickness of ≈ 90m is a tenable estimate from the fragmented successions (Map 4.). At all localities its base is fault bounded, whilst its passage into the overlying mudstones of Penmaen Dewi Formation (Table 4.3) is invariably obscured by tectonism. Fragmented exposures of the Formation crop out at various localities along the coast in the vicinity of Pen Porth-egr [S.M. 8030 3250], north of Aberfelin [S.M. 8330 3275], Pwll Llong [S.M. 8380 3320], and Abercastle [S.M. 8520 3360, Map 4.]. The following section is concerned with the development of the Formation at Pen Porth-egr and Pwll Llong; informally referred to as the Pen Porth-egr member and Pwll Llong beds. The term Pwll Llong beds is essentially a stratigraphic term for a locality where the transition between the Abercastle and Penmaen Dewi Formation can be broadly discerned.

a) Pen Porth-egr member (D1); *Type-section (S.M. 8030 3255)*

The development of the Abercastle Formation on the headland of Pen Porth-egr (hereafter referred to as the Pen Porth-egr member, Table 4.3.) has previously been documented by Cox (1916) and Hughes *et al.* (1982), although the distribution and relationship of the Formation are modified. The close proximity of the Pen Porth-egr member to the Porth-gain Harbour Formation (Map 4.), may suggest that the successions exposed in this region are near to the base of the Abercastle Formation; a suggestion which is supported by their shallow marine character (see below) if the inference of transgressive sedimentation is correct. The member whilst complexly folded appears to show an overall younging to the south, consistent with the stratigraphic development and relations of surrounding strata. To the north of Pen Porth-egr headland the bay of Pen-egr is occupied by highly sheared mudstones from which Hughes *et al.* (1982) record the presence of *Didymograptus sparsus* indicating an Arenig age. It is likely that whilst the section is disrupted it is nearly complete with regard to the development of the Abercastle Formation in this part of SW Dyfed.

Description: Sediments of the Pen Porth-egr member are best exposed on the north facing cliffs at Pen Porth-egr [S.M. 8030 3253]. Here lithologies can be best grouped into *tabular cross-bedded sandstone* which pass southwards through a series of small folds and faults [S.M. 8030 3245] into *coarse sandstone – mudstone alternation* (Fig. 4.4.); although the latter are poorly exposed, or inaccessible.

The *tabular cross-bedded sandstones* are well sorted fine to medium grained and show a wide variety of sedimentary structures. The succession is dominated by tabular and wedge shaped cross-bedded units (Plate 4.8.), although trough cross beds are not uncommon and frequently show overturning of foreset lamination. Tabular cross-sets dip at both low and high angles and in alternating directions, whilst individual laminae are easily traced with lower contacts, which are sharp although non-erosive (Plate 4.8.). Parallel laminated sandstones forming planar beds are also present, within which individual laminae are commonly defined as graded couplets of medium- to fine sandstone varying in size from 1-3mm; more rarely one can observe coarse to fine sandstone couplets. Discrete interbedded wavy and lenticular mudstone partings whilst not common are seen as deeply weathered units varying in thickness from several centimetres to 10's of centimetres.

Overlying the fine tabular cross-bedded sandstones facies, *coarse- to medium sandstone and mudstone alternations* dominated the accessible exposure (Fig. 4.4.). The sequence is comprised of rippled coarse sandstones which show complex bundled cross lamination and irregular lower set boundaries, commonly draped by silt-grade sediment. No bioturbation is observed. The sandstone content diminishes progressively upwards through the sequence towards Pen egr where the marine shales of the Penmaen Dewi Formation are developed; although the transition is complexly deformed. Coarse sandstone to granulestone beds are evident, and documented by Hughes *et al.* (1982) as the basal Arenig conglomerates (their Ogof Hên Formation). They refer however, to the fine tabular cross-bedded sandstones as the Upper Cambrian *Lingula* Flags.

Interpretation; The fine well sorted *tabular cross bedded sandstones* which dominate the lower part of the Pen Porth-egr member are most simply explained as the product of progressive sorting in a marine shoreface. The alternation of low and high angle sets which frequently dip in opposing directions may reflect deposition within the foreshore and upper shoreface (Davis *et al.* 1971). Parallel laminated sandstone are common to such environments where lamination is produced by swash-backwash (Howard 1971), whilst trough cross-beds are developed within the wave breaker-zone (Johnson 1978) which could account for the frequent overturning of forset lamination as seen in the field. The thin discrete mudstone partings are likely to reflect deposition under slightly deeper water conditions outside of the direct influence of wave activity, and may reflect deposition within the lower shoreface.

The passage up into *coarse sandstones and mudstone alternations* is not seen, although the sequence is stratigraphically conformable. The lithologies exhibit features typical of heterolithic clastic sequences (de Raff *et al.* 1977) developed within a shallow marine inner shelf environment, where deposition of sandstones was by wave and storm related events whilst intervening mudstones reflect deposition from suspension during prolonged periods of low-energy conditions. This lithological transition from north to south is thought to reflect deepening of the marine environment (i.e. shore-face – inner marine shelf). Such an interpretation is consistent with a transgressive marine sequence and the spatial attitudes of the lithologies.

Distribution of the Pen Porth-egr member

Elsewhere, the exposed successions of the Abercastle Formation are of similar character to the Pen Porth-egr member. The Formations developed north of Aberfelin [S.M. 8330 3275] and Pwll Olfa [S.M. 8370 3315] are dominated by thick bedded mature sandstones with occasional partings of mudstones and fine laminated siltstones, similar in character to the uppermost successions of the Pen Porth-egr member. Similar lithologies are evident in several of the numerous small quarries around Abercastle, in particular the quarry along the road leading east out of Abercastle towards Garn-isaf [S.M. 8535 3355]; it being probable the Abercastle Formation developed in this area, occupies a similar stratigraphic position. The elongated harbour at Abercastle exposes shales realistically thought to belong to the Penmaen Dewi Formation, although the relations between both Formations can be clearly observed to be faulted near the concrete concourse leading down to the harbour [S.M. 8526 3366]. This locality appears to be that which Cox (1916) and Fortey & Owens (1987) refer to as showing the equivalent of the Porth-gain Harbour Formation as defined herein (their Porth-gain Formation or beds); although there is no evidence of lithologies resembling the polymictic conglomerates and thin wavy and flazer bedded sandstones–mudstone alternations.

(b) Pwll Llong beds (D2). *Type-locality (8410 3325); Thickness (c.20m)*

The term Pwll Llong beds is essentially a stratigraphic term introduced for a thin (*c.* 20m) succession of transitional mudstones and interbedded fine laminated siltstones and sandstones. They are thought to represent the uppermost development of Abercastle Formation and its passage into the overlying Penmaen Dewi Formation; similar transitions are observed elsewhere, although the lithological disparity between both Formations has rendered this horizon an area of inherent weakness during deformation and all contacts are highly deformed.

However, cropping out on the wave-cut platform and cliffs at the south of Pwll Llong [S.M. 8410 3325], the transition between the Abercastle Formation can be broadly discerned. It is reflected by a gradational decrease in the frequency of thin highly convoluted and loaded fine sandstones (?distal storm deposits) as it 'youngs' northwards (Map 4.), to a point where mudstones make up 95% of the exposed lithologies. This transition is interpreted to reflect a deepening in the marine environment to below the effective influence of storm and/or wave activity, with the development of an environment comparable to the outer shelf (Banks 1973). At this point the sequence may be regarded as the Penmaen Dewi Formation; which occupies the majority of the bay at Pwll Llong (Map 4.). At no other point can a similar transition be seen, in this respect the *beds* represent a significant stratigraphic succession.

Summary of Abercastle Formation: The occurrence of *Ogyginus hybridus* within the Abercastle Formation indicates a Whitlandian age (Fortey & Owens 1987). Whilst fragmented it is possible to establish that the Formation concomitantly fines-up from mature sandstones into heterolithic sandstones/mudstones (Pen Porth-egr member) and mudstones (Pwll Llong beds). Such transition is thought to reflect a transgressive marine sequence from shoreface – shallow shelf clastics – shelf muds (Penmaen Dewi Formation); which accords with similar Arenig successions developed elsewhere in SW Dyfed (Table 4.3.). The arenaceous successions of the Trwyn Llwyd Formation (see below) cropping out to the north of Abercastle (Map 4.) are also thought to be of Whitlandian age; it being likely that both Formations represent part of a thick shallow marine clastic complex that developed throughout the area following deposition of the Porth-gain Harbour Formation during early Arenig times.

4.4.3. The Trwyn Llwyd Formation (D). *Thickness (> 400m), 5 members.*

The Trwyn Llwyd Formation represents the most discrete succession defined during this study. The Formation is nowhere exposed in it's entirety and has been pieced together from fault bounded blocks within the coastal successions between Pwl Deri and Pwllstrodur (Map 3.). As outlined {3.4.}, correlation is comparatively well understood due to the development of discrete facies association. More importantly, is the fact that the facies associations are systematically ordered and can be recognised within the allochthonous sequence.

Individual facies association are referred to as Members, the transitions between which are generally gradational reflecting a continuous depositional system (i.e. Pen Deudraeth Member (shelf/prodelta mudstones); Aberbach Member (delta-front sandstones); Godor Mabli Member (heterolithic inner shelf clastics), Ogof Felin Member (shoaling bar sandstones), Ogof Aderyn Member (heterolithic shelf clastic)). Internally however, members are complex, comprising intercalated depositional packages and only broad generalisations are made here. The entire Formation requires detailed study in so much as it offers one of the best preserved siliciclastic shallow marine sequences in the Welsh Caledonides.

The base and top of the Formation is nowhere seen, allowing for diachronism, a thickness of 375-400m is appropriate, although the Formation may well have been thicker prior to deformation. The most complete successions crop out along the coast between the headlands of Aber Mawr [S.M. 8826 3470] and Carreg Golchfa [S.M. 8818 3521]; Godor Mabli [S.M. 8835 3651] and Trwyn Llwyd [S.M. 8785 3662]; Aber Cerrig-gwynion [S.M. 8776 3713] and Penbwchdy [S.M. 8770 3723].

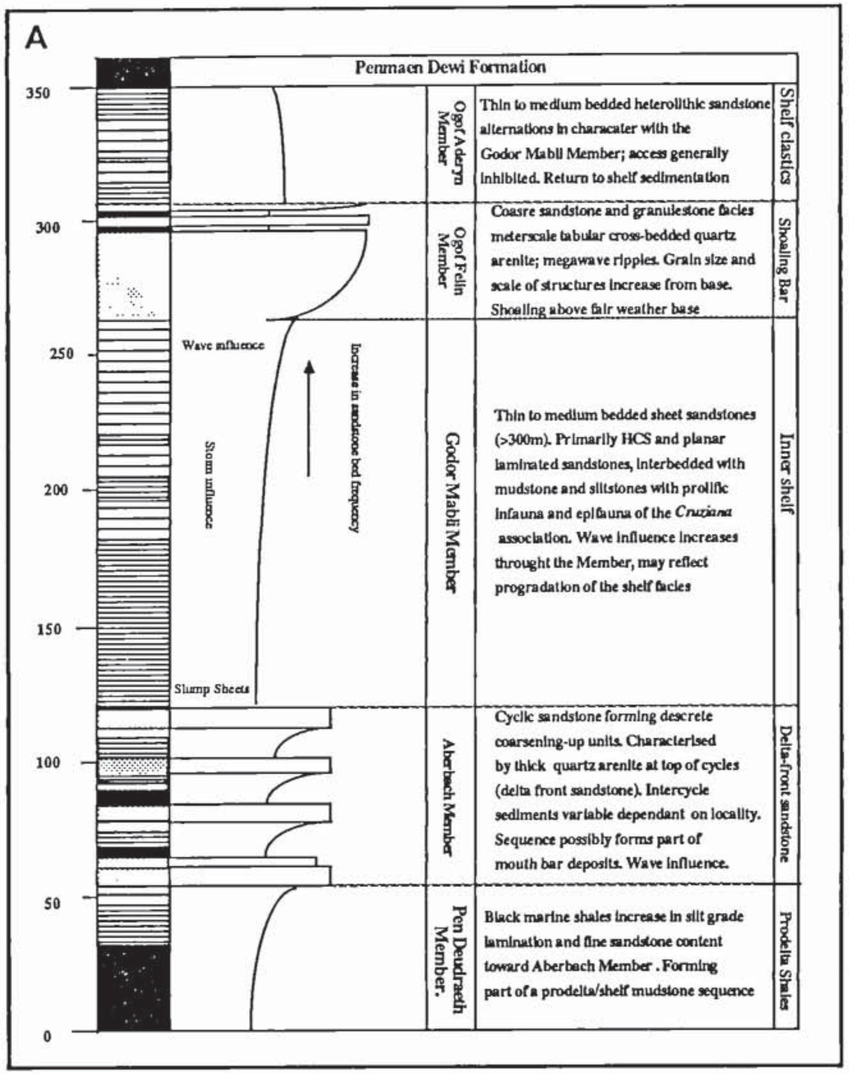
Age of the Trwyn Llwyd Formation

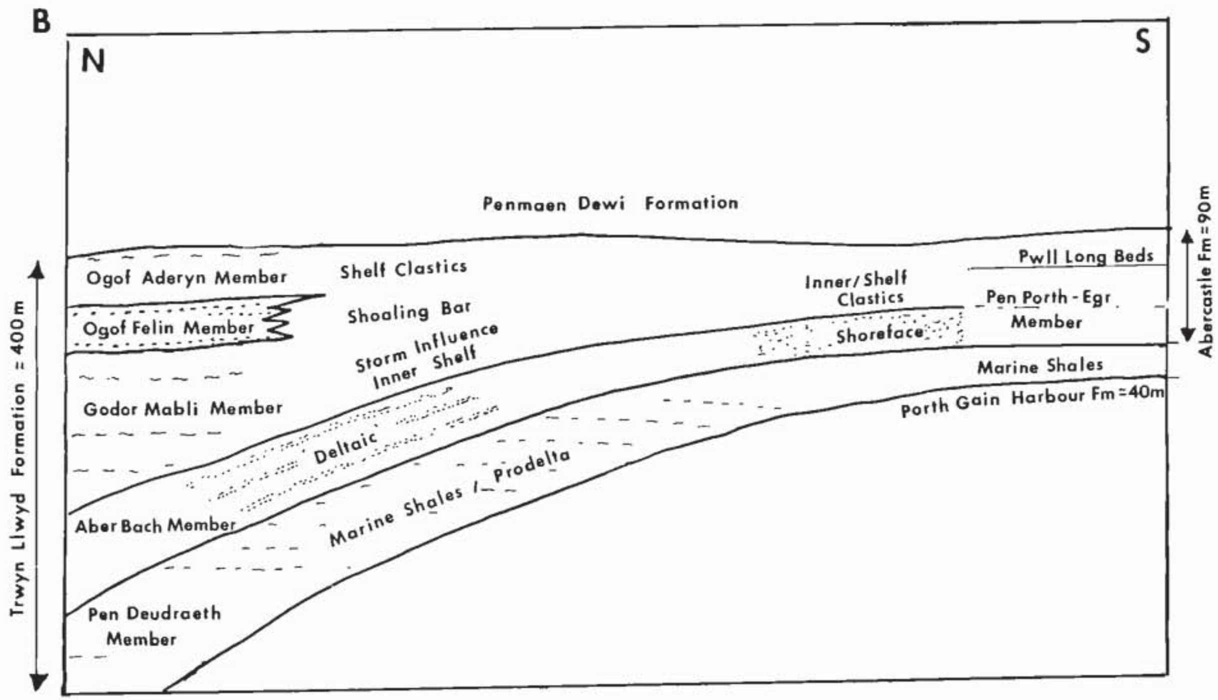
To date, no diagnostic faunas have been collected from the Trwyn Llwyd Formation, although *Callograptus* s.p. is recorded from shales referred to here as the Pen Deudraeth Member (Cox 1930). Dendriod graptolites, including *Callograptus* s.p., are typical of the Whitlandian across much of South Wales (Fortey & Owens 1987), although particular species may have lower ranges. The presence of *Callograptus* s.p. confirms an Ordovician age, which is highly appropriate in the light of the salient similarities that exist between Upper Cambrian sediments of the *Lingula* Group and the Godor Mabli/Ogof Aderyn Members.

Lithologically, the Trwyn Llwyd Formation is unlikely to be of uppermost Whitlandian-Fennian or Llanvirn age, as mudstone deposition persisted during this period (i.e. Arenig-Llanvirn Shales, Fig. 4.1.). The siliciclastic character of the Formation may suggest correlation with the Ogof Hên and Porth-gain Harbour Formations (this study), although the nature of sedimentation with these lowermost Arenig successions is one of rapid transgressive deposition from conglomerates and sandstones into mudstones and siltstones, rather than evolution into a siliciclastic dominated sequence. It is likely therefore, that the Trwyn Llwyd Formation is of lowermost Whitlandian age, and if correct, almost certainly correlates in part with the Abercastle Formation.

However, whilst a lowermost Whitlandian age is thought appropriate, the shales of the Pen Deudraeth Member may relate in part to the Shale member of the Porth-gain Harbour Formation as they probably occupy a similar stratigraphic position. If both sequences are coeval, then the lower parts of the Trwyn Llwyd Formation may span both the Moridunian and Whitlandian; although faunas upon which such an inference can be substantiated, are, at

Figure 4.5.a&b. (**A**) Schematic sedimentary log of the Trwyn Llwyd Formation compiled from several localities between Aber Mawr [S.M. 8826 3470] and Trwyn Llwyd [S.M. 8785 3662]. (**B**) Possible sedimentary and stratigraphic relationships within the Lower/Middle Arenig sequences of the Porth-gain Harbour, Abercastle, and Trwyn Llwyd Formations of the Fishguard – Porth-gain district.


A
Penmaen Dewi Formation
350
300
250
200
150
100
50
0
Wave influence
Storm influence
Increase in sandstone bed frequency
Slump Sheets
Ogof Aderyn Member
Thin to medium bedded heterolithic sandstone alternations in characater with the Godor Mabli Member; access generally inhibited. Return to shelf sedimentation
Shelf clastics
Ogof Felin Member
Coasre sandstone and granulestone facies meterscale tabular cross-bedded quartz arenite; megawave ripples. Grain size and scale of structures increase from base. Shoaling above fair weather base
Shoaling Bar
Godor Mabli Member
Thin to medium bedded sheet sandstones (>300m). Primarily HCS and planar laminated sandstones, interbedded with mudstone and siltstones with prolific infauna and epifauna of the Cruziana association. Wave influence increases throught the Member, may reflect progradation of the shelf facies
Inner shelf
Aberbach Member
Cyclic sandstone forming descrete coarsening-up units. Characterised by thick quartz arenite at top of cycles (delta front sandstone). Intercycle sediments variable dependant on locality. Sequence possibly forms part of mouth bar deposits. Wave influence.
Delta-front sandstone
Pen Deudraeth Member.
Black marine shales increase in silt grade lamination and fine sandstone content toward Aberbach Member . Forming part of a prodelta/shelf mudstone sequence
Prodelta Shales


B
N
S
Penmaen Dewi Formation
Trwyn Llwyd Formation = 400 m
Ogof Aderyn Member
Shelf Clastics
Ogof Felin Member
Shoaling Bar
Storm Influence Inner Shelf
Godor Mabli Member
Deltaic
Aber Bach Member
Marine Shales / Prodelta
Pen Deudraeth Member
Inner/Shelf Clastics
Shoreface
Pwll Long Beds
Pen Porth - Egr Member
Marine Shales
Porth Gain Harbour Fm = 40m
Abercastle Fm = 90m

present lacking.

Concerning the development of the Moridunian and Whitlandian in the immediate vicinity of Fishguard, it is of interest to note that Kennedy (1986) records *Porterfieldia punctata* from Trehail Quarry [S.M. 8830 2880], near Tremaenhir. He suggests that this fauna may represent the development of the *Merlinia rhyakos* Biozone, the uppermost Biozone of the Moridunian. The sediments at Trehail Quarry comprise laminated coarse sandstone and mudstone alternations in character with parts of the Trwyn Llwyd Formation. It seems probable that comparable strata occupies considerable areas of this part of SW Dyfed, much of which is referred to as the *Lingula* Flags {4.3.}.

(a). Pen Deudraeth Member **(D1).** *Type Locality (S.M. 8825 3490) (>40-70m)*

The Pen Deudraeth Member represents a marine shale sequence and forms the base of the Trwyn Llwyd Formation. An appropriate type locality is taken at the small glacial drift covered cliffs north of Aber Mawr and south of the Pen Deudraeth Headland [S.M. 8825 3490]. Here the Member can be seen clearly to pass upwards into the overlying arenaceous Aberbach Member (Plate 4.9.). The exposed thickness of the Member at this locality is approximately 45m, although there is little doubt that its thickness exceeds the outcrop exposure. The base of the Member is nowhere seen. The development of the Pen Deudraeth Member elsewhere, can only be made by inference (i.e. Pwlchrochan [S.M. 8855 3640], Pwlldawnau [S.M. 8803 3710]), although the association and spatial attitude at the above localities with characteristic strata of the Aberbach Member, makes the inference realistic.

Description and interpretation of lithology: The typical lithology of the Pen Deudraeth Member is one of homogenous fissile black shales within which sedimentary structures are uncommon whilst changes in grain size are subtle. Where the transition from the Pen Deudraeth Member to the Aberbach Member is seen, it is apparent that no major hiatus occurred. The transition is characterised by an upward increase in silt and fine grained sand (Plate 4.9.), probably reflecting a shallowing of the marine environment. Fine parallel silt-grade laminae, defined by slight differences in texture and colour, probably reflect deposition from fluctuations in the sedimentation rate and differential settling. Thin, laterally persistent and weakly graded fine sandstones may represent distal storm beds (Johnson 1978, Pedersen 1985). On the wave-cut platform to the south of Pwlchrochan [S.M. 8855 3640] the Member is dominated by marine shales and thin bedded fine grained sandstones containing features typical of current deposition. Individual sandstone beds are variable in thickness (from ≈2-10cm) and show lateral continuity through outcrop. Lower surfaces frequently show loading, above which convolution of lamination is common. Internally, beds may show planar lamination and climbing ripple lamination, grossly resembling turbidites. The suggestion that the sandstones are deposited by currents is supported by the presence of abundant flute marks on adjacent cliff surfaces.

The Pen Deudraeth Member is interpreted as a representing prodelta/shelf mudstone sequence. This interpretation is not based on any one depositional feature or characteristic, rather it is the association with the overlying Aberbach Member (see below). This association indicates that the mudstones can only be interpreted in terms of accumulation on a mud-dominated marine shelf. The fine grained nature of the sediments suggests that sedimentation, was in the main, from suspension, although current deposition periodically transported fine sand grade sediment.

(B). Aberbach Member (D2): *Type Localities (S.M. 8825 3490] and (S.M. 8819 3660].*

The Aberbach Member comprises a series of sandstones, siltstones and mudstones. These form a series of coarsening-up cycles topped by laterally extensive mature sheet sandstones. This feature makes this Member widely recognised in many coastal sections (compare Plates 3.5.a+b. and 4.9.). The Member is best observed at three localities: *a)* Pen Deudraeth and Carreg Golchfa headlands, *b)* north of Pwlchrochan [S.M. 8855 3651], *c)* the cliffs 120m to the west of Godor Mabli [S.M. 8819 3660]. At all of these localities, the distinctive sheet sandstones are clearly evident, although inter-cycle sediments may vary, reflecting the diachronous nature of deposition. Coarsening-up cyclic units are a feature most characteristic of deltaic sedimentation and progradational linear clastic shorelines (see Leeder 1982). The presence of slump sheets (Plate 3.1.), thick homogenous sheet sandstones above coarsening-up cyclic units (Plate 4.9.; delta-front sandstones), and possible prodelta deposits, suggest that the sequence can be most simply explained in terms of 'deltaic' sedimentation. However, the small area with which this present study is concerned, the limited three dimensional exposure, and the complexity of deltaic depositional systems does not allow for a more detailed interpretation.

Description and interpretation of lithologies Cycles within the Aberbach Member are variable in thickness (7-12m) and internally the nature of inter-cycle sediments is often complex. Classic coarsening-up cycles are absent, although the distinctive sheet sandstones are readily recognised. On the Pen Deudraeth headland [S.M. 8825 3490] the Aberbach Member is in excess of 60m thick, where passage upwards from the underlying mudstones of the Pen Deudraeth Member is clearly evident. The lowermost sheet sandstone at this locality is followed by at least four recognisable cycles. The sheet sandstones can be traced from the Pen Deudraeth headland to the opposing cliffs of Carreg Golchfa where the uppermost two cycles

Plate. 4.9. The Pen Deudraeth (D) and Aberbach Members exposed at the south of the Pen Deudraeth Headland (S.M. 8825 3490). The transition is gradational and interpreted to reflect shallowing of the marine environment from shelf mudstones (prodelta) into a delta-front sandstones. The discrete homogenous sheet sandstones characteristic of the Aberbach Member (S.S.) can be traced throughout the coastal districts greatly aiding the stratigraphic interpretation of the area. The base of the sheet sandstone has extremely well preserved linguiod ripples along its base. Arrowed above the sheet sandstone are thick channeled sandstones which are thought to reflect reworking of the bar sandstone. Such channels are not evident in successions further to the north (*n.b.* The Pen Deudraeth Member indicates that thick black marine shales where being deposited prior to the development of shallow marine clastic deposition during Whitlandian times).

Plate 4.10. *Cruziana* from the Godor Mabli Member. Note the preservation of genal spine groves along the edge of the trial whilst V-shaped markings indicate that the trilobite that made this trail was moving from left to right (towards the pubs at Fishguard). Indeterminate inarticulate trilobite fauna have been collected from the associated shales, although further work may as yet find determinate species which would greatly aid the stratigraphic interpretation of the area.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are seen. The overlying Godor Mabli Member is recognisable at the west of Carreg Golchfa and cliffs to the north of the headland, although here both the Aberbach and Godor Mabli Members are complexly deformed (Plate 3.2.).

The nature of cycles at Aberbach are internally complex. The lowermost cycle (Plate 4.9.) passes upwards from the Pen Deudraeth Member without hiatus into a texturally mature sheet sandstone. The base of the sandstone is sharp and features well preserved linguiod ripples. The sandstone is texturally and compositionally mature, reflecting extensive reworking. Weak cross-lamination is locally evident defined by small variations in grain size. The majority of the sandstone is however, structureless and extremely well sorted. This sandstone and all other sheet sandstones which cap individual cycles are most simply interpreted as 'delta front sandstones'. They are thought to form the top a of delta-front sequence possibly within a fluviomarine succession analogous to delta mouthbar sandstones which reflect fluvial and wave interaction (Elliot 1978). There is little direct evidence for tidal influence in the sequence. The maturity of the sandstone in such environments reflects the direct consequence of shoaling at the seaward terminus of the distributory-mouth channel and wave reworking beyond the channel mouth (Coleman & Prior 1980). Above the lowermost sheet sandstone on the Pen Deudraeth headland, a sequence of stacked channels are evident (Plate 4.9.). Such channels may possibly reflect reworking of the delta front sandstone by subsequent fluvial channels as progradation continued seawards.

In the cliffs (≈120m) to the west of Godor Mabli [S.M. 8819 3660] the Aberbach Member is again well exposed (Plate 3.5.b.). The nature of the Member at this locality is one of a lower sandstone:siltstone ratio strongly reflecting the diachronous nature of sedimentation within the deltaic system. Inter-cycle sandstones are absent, with each cycle being defined by a diffuse coarsening-up sequence, terminated abruptly by thick sheet sandstones. The base of each cycle is characterised by black marine shales and thin diffuse silt grade lamination. Bioturbation, whilst evident is not common and individual silt laminae can be easily traced. Thin bedded fine sandstones increase in frequency towards the top of each cycle (Plate 3.5.b.). The lack of intercycle sandstone may suggest that the sequence at this point represents the delta-front situated away from fluvial influence. The sequence may therefore reflect deposition on the lower delta plain that extends seawards from the area actively receiving riverborne sediment, which is characterised by deposits associated with distributory-mouth bars, distal bars, prodelta clays and offshore slump deposits (Coleman & Prior 1980). The Member at this locality contains slump sheets (Plate 3.1.) and sheet sandstones which can interpreted in terms of mouth bar sandstones (Plate 3.5.b.). The presence of silt laminated dark black marine shales which pass upwards into thin alternating fine sandstones may reflect prodelta and distal bar deposition respectively (Plate 3.5.b.). The sequence is comparable to documented delta-front coarsening-upward sequences of the Rhone delta (Oomkens 1967). Here progradation is achieved by beach-ridge accretion and mouth bar progradation, and has distinct similarities to distal mouth bars documented by Johnson & Stewart (1984) from the Brent oilfields of the North Sea. With regard to the latter study, the relationship of the shore/beach complex situated to the landward side of a deltaic system forming a coastal/deltaic clastic wedge, may be appropriate to the shoreface sandstones of the Abercastle Formation and deltaic sandstones of the Aberbach Member, if both sequences are contemporaneous (Fig. 4.5.b.).

Passage from the Aberbach Member into the thick shallow marine siliciclastic of the Godor Mabli Member (Fig. 4.5.) is gradational. There is no apparent brake in sedimentation, although delta-front sandstones are not developed. The transition is observed north of Godor Mabli, where large slump units overly the last recognisable 'delta-front unit'. The increasing frequency of planar laminated and hummocky cross-stratified sandstone, reflect the abandonment of deltation and the onset of deposition on a shallow siliciclastic shelf where sandstone deposition was by wave and storm activity. The reasons for the change may reflect a shift in the site deltation (avulsion) removing the river mouth from the area (Chafetz 1982) or a sufficient rise in sea-level. The nature of the overlying Godor Mabli Member suggests that the latter is the case, with transgression of a shallow siliciclastic sea and delta abandonment taking place in this area.

(c). Godor Mabli Member (D3). *Type Locality (S.M. 8826 3656 to 8896 3664); Thickness (60 m).*

The Godor Mabli Member represents a thick succession of interbedded siltstones and sandstones deposited in a shallow marine shelf environment. They directly overly the deltaic sandstones of the Aberbach Member. The Member has been particularly prone to deformation (Plate 3.2.) being spectacularly folded and faulted at numerous localities along the coast between the headlands of Carreg Golchfa and Aber Carrig-gwynion (Map. 3). It should be noted however, that whilst successions are readily equated along the coastal section due to the relationship with the Aberbach Member, without this relationship, inland exposure of the Member would be relatively indistinguishable (in the absence of faunal evidence) from *Lingula* Group sediments. This highlights a major problem in SW Dyfed regional stratigraphy, in so much as large tracts of the countryside are documented as '*Lingula* Flags', on the basis of alternating siltstones and sandstones (Cox 1916; Williams 1934; Rushton 1974). It seems likely given the thickness of the Trwyn Llwyd Formation that some of the '*Lingula* Flags' successions documented north of St. Davids are in fact of Arenig age.

West of Godor Mabli (between Godor Mabli and Trwyn Llwyd [S.M. 8826 3656 to S.M. 8896 3664]) the Godor Mabli Member appears to be complete, and serves as an excellent type section. It illustrates passage from the underlying Aberbach Member and the transition into overlying coarse sandstones of the Ogof Felin Member on the Trwyn Llwyd headland (Map 3.). Minor faults are observed in the section although little of the Member is missing, its thickness here is in excess of 160m.

Description of and interpretation of lithologies: The typical lithology of the Godor Mabli Member, is one of thin-medium bedded siltstones, fine-medium sandstones, and rare mudstones. The finer sediments generally weather grey to buff, the sandstones weather buff white to light brown-fawn, with the variation in colour reflecting relative maturity and sandstone content. The discrepancy in grain size and colour sharply defines bedding. Sedimentary structures on the whole are well preserved, except where bioturbation has caused homogenisation; although trace fossils are often well preserved. At its type section [S.M. 8826 3656 to S.M. 8896 3664], the Member shows an overall tendency to coarsen upwards. This large scale mineralogical change whilst obscure when viewed directly at outcrop, can be seen from the south on the opposing headland of Carreg Herefio looking northwards across Pwlchrochan. Here a gradational colour change can be seen from dull grey in the east above the Aberbach Member, to light-grey and fawn/light brown sediments to the west below the Ogof Felin Member on the Trwyn Llwyd headland.

The Godor Member can be best described in terms of a *siltstone-mudstone association* (Fair-weather sedimentation) and *sheet sandstone association,* variations on which are prevalent, reflecting the variable hydrodynamic conditions that prevailed during deposition. Both lithologies can be regarded as 'end-member' facies.

The *sheet sandstone association* consists of thin to medium and more rarely thickly bedded units, that form laterally continuous sandstones traceable over extensive distances (up to 300m). Local discontinuity is evident, upper surfaces commonly swell and pinch giving an undulatory appearance to many beds. The grain size of such units is predominantly fine to medium, coarse sands are rare although occasionally evident at the base of some thicker units. The most commonly observed sedimentary structures are parallel to low angle planar and tabular cross lamination, and hummocky cross-stratification (Plate 4.11.). Hummocky cross-stratification is commonly regarded as a storm related bed form (see Duke 1984, and therein) characteristic of storm influenced inner marine shelf sedimentation. Such stratification is sufficiently common to suggest that the Godor Mabli Member reflects a storm influenced inner shelf sequence.

Mudstones and siltstones represent periods of prolonged fair-weather conditions, during which times deposition was primarily by suspension and minor current related processes. The facies is characterised by fine siltstone to mudstone grade sediment, within which sedimentary structures are commonly homogenised by bioturbation. The nature of biogenic structures is highly variable and well preserved, and almost certainly that of the *Cruziana* association (Seilacher 1967, 1978). Bedding surface trails are prolific where the effects of burrowing are minimal. The nature of trails is dominated by *Cruziana* and *Rusophycus*, preservation is significantly good with small scale features such as lateral striations, and genial spine indentation grooves preserved (Plate 4.10.). Burrows are similarly well preserved, and resemble in the most part the sub-horizontal to inclined burrows of *Teichichnus* and gently curved single burrows of *Planolites*. Collectively, the traces are thought to reflect the activities of a diverse benthic community of mobile epifaunal and shallow burrowing infaunal detritus feeders, which colonisied a moderately stable, well oxygenated substratum characteristic of the inner shelf (Mount 1982). Where homogenisation is limited, the siltstones and mudstones show a variety of structures. The most common structure is parallel lamination of siltstone and fine-medium grained sandstone, the latter may show evidence of a grading.

The Godor Mabli Member is thought to reflect deposition on a shallow marine storm influenced inner shelf (*Cruziana* association), where sedimentation was by storm and fair weather processes, comparable successions having been documented by de Raff *et al.* (1977), Brenchly & Newall (1982) and Soegaard & Eriksson (1985). The overall coarsening-up of the Godor Mabli Member may reflect progradation of the shelf facies, although an increase in the coarse sediment supply may be equally as important (Reinech & Aieger 1982). In summary, mudstones and siltstones were deposited during periods where fair-weather conditions prevailed. Small discontinuous and graded sandstone lamination may reflect low intensity storm agitation. Extensive bioturbation during periods of fair-weather conditions homogenised much of the fine sediment. Sheet sandstones were periodically pulsed into the environment by wave and storm related processes in catastrophic events (note escape burrows in hummocky cross-stratified sandstones in Plate 4.11).

(D). Ogof Felin Member **(D4).** *Type Locality (S.M. 8785 3665); Thickness (c. 40m).*

The Ogof Felin Member is a succession of massive bedded medium to coarse sandstones and granulestones which conformably overlie the Godor Mabli Member. This sandstone dominated Member is atypical of other members due to its thickness and maturity which makes the succession readily identifiable at outcrop. The most complete section (*c*. 40m) is exposed on the headland of Trwyn Llwyd [S.M. 8785 3665], where it can be clearly seen that the Member passes from the underlying Godor Mabli Member upwards and conformably into the overlying Ogof Aderyn Member. The Ogof Felin Member outcrops on the headland of Carrig Gwynion [S.M. 8780 3730, Map 3]. Here the member is again readily identifiable due to its thick sandstone geometry and clastic maturity. The thickness of the Member at Carrig Gwynion is less than 25m, although it is clearly evident that it represents an anomalous sandstone facies

Plate 4.11. Hummocky cross-stratified sheet sandstone from Godor Mabli Member. Note vertical escape burrows (arrowed). Such sandstones are typical of storm influenced inner shelf sequences where deposition is commonly by periodic catastrophic events.

Plate 4.12. Meter scale tabular cross-bedded texturally mature sandstones of Ogof Felin Member [S.M. 8785 3665]. It is thought that such large structures may relate to the sand wave/proximal slope subfacies of an off-shore bar (see Johnson 1978). Note low angle discordance of erosion surface which the hammer is resting on (hanging-on). At outcrop these beds are vertically dipping and the plate has been rotated through 90° from left to right to give the depositional attitude. Plate 4.13. shows large wave rippled sandstones taken from the same member below this sequence and may reflect the initial sand wave facies.


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within heterolithic shallow marine sandstones and mudstones of the Godor Mabli and Ogof Aderyn Members which occupy the area north and south of the succession (Map 3.).

Description and interpretation of lithologies; Passage from sediments of the Godor Mabli Member to the Ogof Felin Member on the headland of Trwyn Llwyd is marked by an increase in the frequency and thickness of sandstone beds. The 'fair-weather' siltstones and mudstones of the Godor Mabli are lost and the sequence is composed entirely of sandstone (Fig. 4.5.). Prior to the transition, hummocky cross-stratified units and cross-laminated sandstone with wave rippled tops dominate the sequence. Bioturbation becomes less frequent (in comparison with the lower parts of the Godor Mabli Member) to a point where it is absent. This is interpreted as reflecting a higher energy state of the depositional environment and less favourable conditions for marine fauna. It represents depositional conditions, where, if mud was allowed to accumulate it was rapidly reworked, effectively reflecting shoaling of the environment to above fair-weather base. The top of the Member as seen at the north of the Trwyn Llwyd headland [S.M. 8795 3670], grades upwards into the overlying Ogof Aderyn Member (Fig. 4.5.). The transition records an abrupt increase in the frequency of interbedded mudstones, with the return to heterolithic lithologies.

Internally, the Ogof Felin Member shows a wide variety of sedimentary structures. An overall increase in grain size is evident, with a higher proportion of medium to coarse grained sandstones at the base, passing upward into predominantly coarse/very coarse grained sandstones and granulestones. The scale of sedimentary structures also increases up section, from bidirectional tabular cross-bedded sandstones through large scale wave rippled sandstones (Plate 4.13.) into meterscale tabular cross-bedded sandstones (Plate 4.12). From one exposure, such features may relate to a number of subfacies with any one particular shallow marine 'shoaling' facies (see Davis 1983). The reason for suggesting that the Member represents part of an open marine shoaling bar is based on three features: 1) the transition from shelf processes influenced by storm/wave and fair-weather sedimentation into a coarse facies which was wave dominated (Plate 4.12. 4.13.) reflecting a shallowing of the marine environment to above fair-weather wave base; 2) structurally the member shows gross similarities with documented offshore bars in an overall coarsening-up in grain size and an increase in the geometric scale of sedimentary structures (Johnson 1978 and therein); 3) the Member is hosted within siliciclastic shelf sediments (Fig. 4.5.), supporting the view that the Member reflects a transient shoaling facies.

Present day storm dominated shelves characteristically contain arrays of elongated sand ridges (e.g. Swift 1978), whilst analogous ancient successions are common in the literature (e.g. Cotter 1985). It seems likely, given the geometry of the Ogof Felin Member and its position between shallow shelf clastics (i.e. Godor Mabli & Ogof Aderyn Members (Fig. 4.5.)) that the succession is comparable with documented bar complexes and reflects prolonged shoaling of the marine environment above fair weather base.

(e). The Ogof Aderyn Member (D5). *Type Locality [S.M. 8795 3670] (Thickness > 70m)*

The Ogof Aderyn Member conformably overlies the Ogof Felin Member to the north of the Trwyn Llwyd headland [S.M. 8795 3670]. The transition between both members is represented by the occurrence of wavy and lenticular bedded mudstones and siltstones interbedded with coarse sandstones towards the top of the Ogof Felin Member. This lithological change is thought to represent a gradual abandonment of shoaling (probably related to increased depth and/or lack of sand supply) with a return to shallow marine shelf sedimentation and the conditions which existed during the development of the Godor Mabli Member.

Access at outcrop to the majority of the Ogof Aderyn Member is not possible. However, the gross features of the Member (sandstone-mudstone alternations) are clearly seen to be similar in lithological character to those of the Godor Mabli Member. The largest exposed thickness of he Member is in excess of 70m. A further succession of interbedded shallow marine clastics equated to the Ogof Aderyn Member is seen in Aber Cerrig-gwynion bay [S.M. 8780 3730], cropping out north of the Ogof Felin Member as developed on the headland of Cerrig gwynion (Map 3.). The Member here is cut by the large tonalite intrusion which occupies the headland of Penbwchdy, the contact with which is non-structural as seen by a thin zone of silicification and large (30cm-1m) sandstone and silt xenoclasts incorporated into the intrusive body. To the north of the intrusion (Map 2.), a further succession of sandstones and laminated mudstones equated with the Ogof Aderyn Member crop out in a small tract to the south of Pwll Deri [S.M. 8875 3790]. At this locality, sediments are complexly folded whilst exposure is generally poor, although the alternating heterolithic character of the Member is still clearly recognisable. *Thallasonodies* and other trace fossils are spectacularly preserved along bedding planes in exposures near the tonalite contact [S.M. 8860 3805]. The Member at this point is overlain by shales from which Cox (1930) records the presence of extensiform graptolites indicating an Arenig age (i.e. Penmaen Dewi Formation (Map 2.)). Whilst the contact is not seen, it seems likely that the development of the Ogof Aderyn Member in this area is equivalent to the top of the Trwyn Llwyd Formation.

Correlation of the Trwyn Llwyd Formation

The Trwyn Llwyd Formation appears to extend significantly further north towards the Prescelly Hills. Here Lowman (1977) documents a fault bounded sequence of Arenig clastic sediments (his Trehallen Beds,) which show evidence for cyclic sedimentation. He suggests that they may reflect delta-building, although exposure is poor within the inland sequence. Elsewhere, no strata resembling the Trwyn Llwyd Formation has been identified, although as outlined previously, much of the SW Dyfed is documented as *Lingula* Flags which show salient features common to the Godor Mabli and Ogof Aderyn Members.

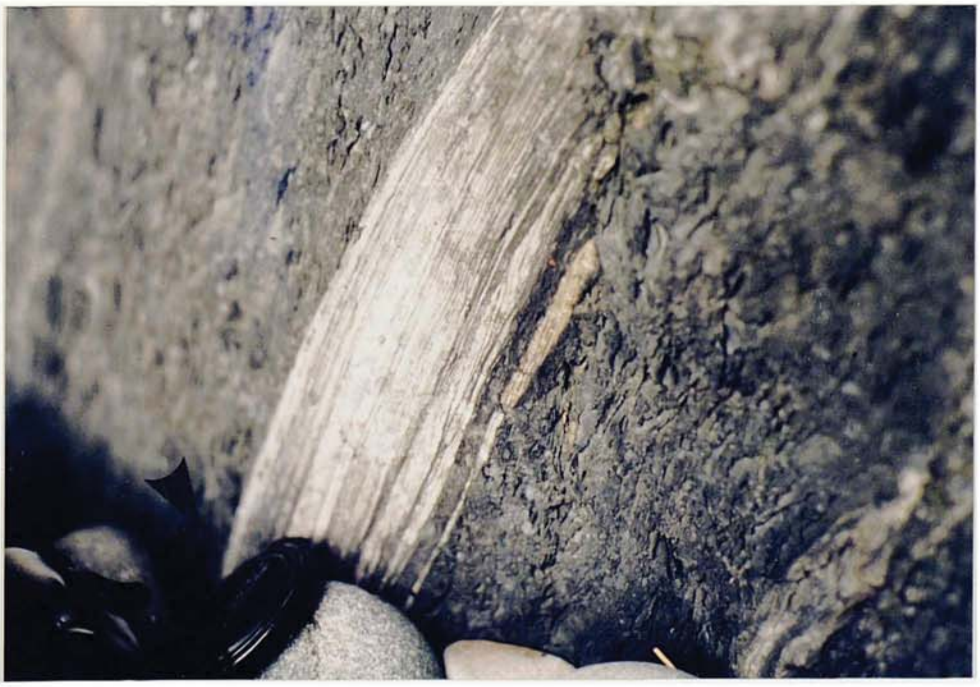
Stratigraphically, both the Abercastle and Trwyn Llwyd Formations are likely to equate with the 'Brunel Beds' of Thomas & Cox (1924) which overly the Trefgarne Volcanic Formation. They contain a graptolite fauna including *D. extensus* Hall, and trilobite fauna which includes *O. hybridus* indicating a Whitlandian age (Thomas & Cox 1924, Fortey & Owens 1987). The Brunel Beds appear to be lithologically similar to the Abercastle Formation, although description of the successions around Trefgarne is limited. Further successions of beds collectively referred to as the Brunel Beds, crop out in the Tremaenhir syncline and surrounding areas (Williams 1934). The exact ages of the various successions are uncertain, with recent faunal evidence suggesting a Moridunian age may be appropriate to part of the sequence (*pers. comm.* R.J. Kennedy 1987). Further to the east, the Blaencediw and Colomendy Formation's are likely to be a regional stratigraphic equivalent (Table 4.3.), although they have been interpreted as mass-flow deposits and turbidites (Fortey & Owens 1987) and may represent a deeper marine facies. On Ramsey Island, Kokelaar *et al.* (1985) whilst identifying the lower Moridunian and Fennian (Table 4.3.), have not identified Whitlandian sediments. They attribute their absence to syn-volcanic sedimentary sliding resulting in omission of the Arenig stratigraphy. Whilst this may be a possible cause, an alternative explanation is that the Whitlandian sediments may not have developed in any great thickness in this area. This is supported by the presence of shallow marine sediments of the Trwyn Llwyd and Abercastle Formations north of Ramsey Island (i.e. basin-wards in the classic Welsh Basin model), therefore this area may have represented a topographic high or very shallow marine sequence during the Whitlandian.

Summary of environments of deposition

The Trwyn Llwyd Formation represents a thick accumulation of predominantly clastic sediments, deposited for the most part, above storm wave base. The sedimentary environments that prevailed during deposition of the Formation include, a storm influenced shallow marine platform (i.e. Godor Mabli Member and Carreg Gwynion Member) that suffered periodic shoaling which resulted in the development sand bodies (i.e. Ogof felin Member). This sequence transgressed sediments deposited in an environment analogous to a subaqueous delta-front facies (i.e. Pen Deudraeth Member and Aberbach Member). Such an interpretation of depositional environments within the mid-Arenig of SW Dyfed, indicates that the area was a

Plate 4.13. Large scale symmetric wave ripples within the Ogof Aderyn Member. The textural maturity of the sediment and common occurrence of wave generated structures reflects progressive sorting above fair-weather base and shallowing of the marine environment from the argillic/siliciclastic sequence of the underlying Godor Mabli Member.

Plate 4.14. *Silt turbidite.* Fine silt-laminated turbidite from marine mudstones at Dinas Mawr [8870 3875]. The age of the sediment is uncertain, although similar turbidites are common to the Arenig-Llanvirn Shales. Note isolated lenticular ripple at base (fading ripple) passing through a thin mudstone layer into climbing ripple lamination. Lens cap is arrowed.



complex depositional centre hitherto unappreciated. Possible relationships between the Trwyn Llwyd Formation in the north and the Porth-gain Harbour and Abercastle Formations to the south, are depicted in Figure (4.5.b.). The figure speculatively takes into consideration both the stratigraphy, thickness variation, and observed sedimentology.

4.5. MIDDLE ARENIG – UPPER LLANVIRN

Deposition of arenaceous material appears to have abated throughout much of SW Dyfed towards the middle part of the Whitlandian (Fig. 2.2.). From this time, marine mudstones accumulated within a relatively deep and partially restricted marine environment; conditions that were to persist until the late Llanvirn. During this period of marine 'quiescence', the region was a major site of igneous activity. Evidence for this is indicated by numerous accumulations of subaqueous volcanics and the contemporaneous emplacement of high-level intrusives into the argillic pile. This following section, briefly documents the stratigraphy of the mudstones developed in the Fishguard – Porth-gain district, followed by an outline of the limited sedimentology offered by this relatively monotonous sequence.

4 5.1. The Arenig-Llanvirn Shales

Large areas of the Fishguard – Porth-gain district are occupied by black marine shales; referred to collectively as the 'Arenig-Llanvirn Shales' (Maps 1-4). It is likely that the successions range in age from the mid-Whitlandian to the Upper Llanvirn, although diagnostic faunas are generally absent. In an attempt to locally differentiate the sequence, the relationships with other strata has been considered (Maps 1-4) and the faunal evidence of previous workers assessed. The subdivisions adhered to are those of Fortey & Owens (1987: i.e. Penmaen Dewi Formation, *D. artus* Shales, *D. murchisoni* Shales; Table. 4.1.). The cumulative thickness of the mudstone sequence is extremely difficult to establish given the extensive tectonism and lack of biostratigraphic control. At best an estimate of ≈300m may be appropriate, although this is not based on any one section; whilst the presence of the *D. artus* Shales appears as yet to be proven (see below).

Following an inference initiated by Cox (1930) that the shales which underlie the Porth Maenmelyn Volcanic Formation [S.M. 8880 3925; Map 5] resemble the Lower Llanvirn of the stratotype sequence east at Traeth Llfyn and Abereiddi Bay (see Hughes *et al.* 1982), it has be customary to denote all sediments underlying the Fishguard Volcanic Group as belonging to the *D. artus* Shales. This has led to the widely held view that the Fishguard Volcanic Complex is of Lower Llanvirn age. However, whilst black shales crop out sporodically below the Volcanic Complex (Maps 1 & 2), particularly to the south of Fishguard, fauna from shales in direct association with the Volcanic Complex at Lower Town, Fishguard, indicate the presence of the lower *murchisoni* Biozone (Lowman 1977, Kennedy 1986). Therefore it is highly likely that the Fishguard Volcanic Complex is confined to the Upper Llanvirn rather than the Lower Llanvirn {5.4.}.

(a). Penmaen Dewi Formation **(*Tetragraptus* Shales) (F1).** *Thickness (? >130m)*

The Penmaen Dewi Formation, a collective term for mudstones of Arenig age, is developed at several localities. Passage from the Abercastle Formation into the Penmaen Dewi Formation has been discussed (i.e. Pwll Llong beds: [S.M. 8415 3335]). The transition probably reflects a gradational change from thin shelf sandstones which pass into a mudstone dominated shelf facies reflecting a steadily deepening marine environment. The succession at Pwll Llong may be regarded as a representative locality for the development of the lower parts of the Formation in this area. Passage into the Llanvirn has not been identified.

In the absence of evidence to the contrary, it is appropriate to assign shales documented by Cox (1916, 1930) containing extensiform graptolites, to the Penmaen Dewi Formation (Maps 3 & 4). The successions are,

a) Aberfelin [S.M. 8340 3250], poorly preserved extensiform graptolites
b) Pwll Olfa [S.M. 8370 3300], *Didymograptus nitidus* Hall, *Didymograptus* cf. *extensus*
) Pwl Whitting [S.M. 8420 3360], *Didymograptus* cf. *sparsus*
d) Pwll Deri [S.M. 8930 3850], poorly preserved extensiform graptolites

indicating that the majority of the exposed shales in the Porth-gain – Abercastle – Pwll Deri district are of Arenig age. This is supported in part by the close spatial attitude of the shales at the above localities with sediments of the Abercastle Formation and Trwyn Llwyd Formation (Maps 2,3,4.). Elsewhere, Hughes *et al.* (1982) record a poorly preserved Arenig fauna, including *D. sparsus*, from the north-east corner of Pen Porth-egr [S.M. 8040 3240]. The shale succession in this area is fault bounded against the Abercastle Formation (Pen Porth-egr member) and is viewed as being of similar stratigraphic character to successions elsewhere. The thickest exposure of shales equated with the Penmaen Dewi Formation, crop out along the coast between Aber Yw [S.M. 8540 3375] and Pwllstrodur [S.M. 8660 3480], an inference based on their close along strike proximity to shales at Abercastle and Pwll Llong. This is supported by the ability to intermittently trace conspicuous penecontemporanueous autobrecciated felsic sills from Pwll Llong to Pwllstrodur (Maps 3 &4; for details see Chapter 5).

The presence of *D. nitidus* (Hall) and *D. c.f. extensus* may indicate that the shales at Pwll Olfa represent part of the *D. nitidus* sub-biozone an intermediate sub-biozone of the *D. extensus* Biozone. This is in keeping with a Whitlandian age, as the *D. nitidus* sub-biozone appears broadly equivalent. The presence of *D.* cf. *sparsus* at Pwl Whitting and Pen Porth-egr may indicate that the shales in these areas belong to the *D. hirundo* Biozone; broadly equivalent to the Fennian Stage. Recently, *D.* cf. *sparsus* has been collected from shales to the south-east of Fishguard at Panty-y-Phillip Quarry, Scleddau [S.M. 9480 3355] (*pers. comm.* R.J. Kennedy); possibly indicating the widespread development of shales of Fennian age.

(b). *D. artus* Shales (*D. bifidus* Shales, Lower Llanvirn) (F2).

The Lower Llanvirn *D. artus* Shales, previously referred to as the '*Bifidus* Shales', have long been suggested to crop out within the Fishguard area. However, only at Aberfelin [S.M. 8340 3250] has an assemblage which suggests a Lower Llanvirn age been identified (Cox 1916; i.e. *D.* cf. *bifidus* (Hall), *D. stabilis* Elles & Wood, and *Bellerophon* sp.). During this study, and that recently completed by Kennedy (1986) who made an extensive faunal search throughout Fishguard district; no diagnostic Lower Llanvirn faunas were found. It would appear that the identification of the *artus* Biozone east of Abercastle has, as yet, to be proven. Excluding the stratotype Llanvirn at Abereiddi Bay (see Hughes *et al.* 1982), the nearest diagnostic *artus* assemblage is recorded from Ty-newydd Grug (Kennedy 1986), some 10kms south of Fishguard. However, Lowman & Bloxam (1981) assign much of the area south of Fishguard to the*'Llanvirn Shales'*. It is likely that the sequences cropping out between Fishguard and Scleddau may contain the Lower Llanvirn, given the development of the Upper Llanvirn at Fishguard (see below) and recorded Arenig fauna from Scleddau.

(c). *D. murchisoni* Shales (Upper Llanvirn) (F3). *Thickness (?>100m)*

The spatial distribution of the Upper Llanvirn remains largely uncertain, despite the presnce of a diagnostic*'murchisoni'* fauna at several localities (Map 1.). At Tower Hill, south of Fishguard [S.M. 9635 3685], Reed (1895) records the presence of an inarticulate trilobite assemblage which he suggests is indicative of the Lower Llanvirn. Recent re-examination of Reed's specimens however, suggests a more restricted assemblage containing several indeterminate fragments (*pers. comm.* R.J. Kennedy). Renewed collecting at Tower Hill has yielded a varied fauna of graptolites, cephalopods, brachiopods and rare trilobites; the graptolites have been identified (*pers. comm.* R.B. Rickards; in Kennedy 1986) as *D. murchisoni* , *D. artus* and *Glyptograptus latus* (Bulman), an assemblage indicative of the lower *murchisoni* Biozone. The *D. murchisoni* Shales at Tower Hill spatially underlie the Fishguard Volcanic Complex in an apparently continuous NE–SW striking belt (Map 1.) and almost certainly equate with shales hosted within the Volcanic Complex, from where Lowman & Bloxam (1981) record a similar lower *murchisoni* Biozone assemblage {5.4.}. The thickness of the *murchisoni* shales at Tower Hill is likely to be in excess of 100m.

Regional Correlation: Regional correlation of the various Arenig to Upper Llanvirn shale successions throughout much of SW Dyfed has recently been addressed by Kennedy (1986, *in press*) and Fortey & Owens (1987). In the absence of faunal control in the Fishguard district, little can be added to the discussions given by the aforementioned workers, and that of Hughes *et al.* (1982) with regard to the development of the type Llanvirn at Abereiddi Bay. It is of interest to note however, that whilst fragmented, the development of the 'argillic' Arenig may be complete within the Fishguard – Prescelly district. The successions between Pwll Llong and Pwl Whiting reflecting the development of the upper Whitlandian and Fennian, comparable to the Road Uchaf Formation on Ramsey Island (Kokelaar *et al.* 1985), and the Pontyfenni and Llanfellteg Formation developed around Carmarthen (Table 2.1.). The development of the Lower Llanvirn appears, as yet, to be proven in the Fishguard district. The Upper Llanvirn *D. murchisoni* Shales are developed south of Fishguard, although local correlation and relationship with other strata remains uncertain.

Sedimentology of the Arenig-Llanvirn Shales

Collectively, the Arenig-Llanvirn Shales comprise monotonous, black, pyritiferous mudstones which afford little in the way of recognisable sedimentary structures or other depositional features. Interpretation of the marine environment during this prolonged period of sedimentation is therefore dificult. The identification of specific 'black shale' facies in ancient successions requires the broad appreciation of a range of disciplines from the palaeotectonic environment and eustatics to the interplay of biogenic and sedimentary processes; it being widely recognised (Greensmith 1978, Arthur *et al.* 1984) that depth of water has little or no controlling influence upon black shale formation. This feature can be demonstrated in this study area by comparing the Pen Deudraeth Member of the Trwyn Llwyd Formation, and the observations presented below.

This following section briefly documents some observations from the mudstones at Aberfelin [S.M. 8335 3250] where the sequence is believed to be of Arenig age (i.e. Penmaen Dewi Formation), and Dinas Mawr [S.M. 8880 3875] where the age is uncertain. Both localities afford sedimentary structures which are sufficiently preserved, so as to allow an insight into depositional processes. Whether or not such processes are applicable to the entire Arenig- Llanvirn Shale sequence is a matter of conjecture.

Description and interpretation of lithologies: The wave-cut platforms at Aberfelin and Dinas Mawr (whilst not strictly representative of the Arenig-Llanvirn Shales in that they afford determinate sedimentary structures) contain laminated siltstone/mudstone sequences and a volumetrically minor, although significant volcanogenic component. Sedimentary structures and bed forms are generally highlighted by colour variations resulting from variable grain size (mudstone to siltstone) and, less frequently, disparity in composition (i.e. tuffs and argillites). Sediments from both localities can be broadly grouped in terms of, *homogenous black mudstones*, *laminated silt-mud turbidites*, and *silt- to fine sand grade volcanogenic mass-flows*; the latter crop out only at Aberfelin and are considered in more detail under the heading of igneous geology {5.2.}. All lithologies are best observed on wave polished surfaces.

Homogenous black mudstones which show little or no apparent structure, other than occasional faint diffuse lamination are common. Individual units, defined by laminated silt-mud turbidites (see below), vary from several centimetres to tens of centimetres within which there is virtually no internal bedding or other primary sedimentary structure. This may reflect biogenic homogenisation, although evidence to indicate the activity of burrowing infauna is generally lacking. This is similarly the case within the interbedded laminated turbidites which would tend to preserve such activity by virtue of their structure, or at least show some evidence of admixture along top surfaces. In the absence of bioturbation, it is therefore thought that the structureless mudstones have an origin comparable with 'pelagic-settling' common to outer marine fans and abyssal plains (e.g. Weering & Iperen 1984); although true deep marine onditions are not envisaged (see below). Interbedded volcanoclastic turbidites and marine shales of the Carreg Member (Map 6) locally show extensive homogenisation by burrowing infauna, whilst much of the Arenig-Llanvirn Shale sequence which show limited evidence for sedimentary structure may reflect intense biogenic activity in character with modern mud-dominated shelf and deeper oceanic sediments (Dojes & Hertweck 1975). It is of interest to note that structureless mudstones have been ascribed to mud-charged turbidity currents ('unifites') and less-dense turbid layer flows (Stanley 1981), whilst faint laminated sequences may reflect disorganised turbidites (Stow & Piper 1984).

Laminated silt-mud turbidites: Whilst 'structureless' mudstones are common, by far the most distinctive and most important feature at Aberfelin and Dinas Mawr, is the occurrence of silt-laminated units. These are recognised locally by a gross fining-up sequence, although more importantly by regular vertical sedimentary structure which can be interpreted in terms of the mud turbidite 'Bouma' signature (i.e. T_0 - T_8 & P = Pelagite/Hemipelagite; see, Piper 1978, Stow & Shanmugam 1980). Complete 'ideal' 9 tiered sequences are absent, whilst the observed structures and ordering of such structures are not communal to every definable turbiditic unit; variability in thickness and type/order of structures being common. Nevertheless, ordered structures are sufficiently preserved (Plate 4.12.) to indicate that deposition was from wanning-current flows. The identification of individual silt turbidites appears important with regard to the local palaeogeography, in so much as thick black marine shales (Pen Deudraeth Member) were developing in relatively shallow marine environments.

At Aberfelin individual turbidites are locally difficult to resolve, due in part to the relatively small percentage of silt grade sediment within the sequence; although such problems may be regarded as typical of fine turbidic muds (Stow & Piper 1984). Where individual turbidites can be readily discerned they characteristically show: lateral continuity through outcrop, variability in thickness from 1-10cm (generally 3-5cm), normal grading with regard to silt content, lower surfaces which are generally sharp to weak undulatory although not scoured, and a diffuse upper surface. Internally, a typical unit can be characterised by thin lower divisions of weakly convoluted and irregular silt laminae with long wavelengths (T_2). These passes into a graded weak planar to non-laminated silty-mudstone (T_3). This in turn may pass upwards into thin indistinct parallel lamination and wispy lamination (T_{4-5}).

At Dinas Mawr, no such problems occur with regard to the definition of individual turbidites. They are discretely defined as buff coloured, silt laminated units within black pyritiferous mudstones which show little or no evidence of lamination (Plate 4.14.). Individual turbidites are generally less than 10cm thick showing lateral continuity through outcrop and containing a wide variety of structures indicative of silt-laminated turbidites. Plate 4.14. shows a typical turbidite within which recognised layering suggest the presence of T_0 to T_6 ubdivisions. In Plate (4.14.) the base of the turbidite (T_0) is recognised by a thin discontinuous lenticular ripple (fading ripple) isolated from the main turbidite unit by a mud layer (T_1), which passes upwards into low amplitude climbing ripples (T_2). This in turn gives way to thin regular parallel silt lamination (T_3) passing into weakly structure silt (T_4) and fine planar laminated silt (T_5). The top of the turbidite is sharp passing abruptly through fine indistinct lamination (T_{5-6}) into mudstone. Collectively, such features are typical of silt laminated top-cut-out turbidites which where deposition was from a waning current (Stow *et al.* 1984).

Silt-grade volcanoclastic mass-flows deposits occur at numerous intervals throughout the Arenig-Llanvirn Shales. At Aberfelin they are readily distinguished due to their vivid bleached colouration, being composed of unresolved silt-grade detritus and fine subangular feldspar and quartz. However, the most characteristic feature of the flows at this locality is the strongly erosive nature of individual flow units, reflecting high-density and high-velocity currents (see Chapter 5 for details). The source of such deposits is uncertain, although they may represent the 'distal-end' of larger 'proximal' debris flows which were generated from slope failure of talus apron accumulation around areas of volcanic activity {5.2.}. Within such areas (as is apparent from the Fishguard Volcanic Complex {5.4.}), during periods of volcanic quiescence the nature of sedimentation reverted rapidly back to mudstone.

Summary and palaeobathymmetry

Deposition of the thick Arenig-Llanvirn Shales in the Fishguard – Porth-gain district, and as developed elsewhere in SW Wales, represent a problematic succession upon which to base a broader palaeogeographic interpretation. There is little doubt that the black shales accumulated as mudstones in an essentially anoxic marine environment into which within coarse terrigenous material had limited access over a prolonged period. Deposition was by pelagic processes, turbidity currents, and associated mass-flows. Thin laminated mudstones possibly represent seasonal cycles, whilst structured finely laminated units represent deposition by muddy turbidites. Volcanogenic silt-grade mass-flow deposits were periodically introduced into the basinal environment as high-particle, high-velocity turbidites, contributing a volumetrically minor, although significant detrital component.

The presence of silt-mud turbidites and carbonaceous marine shales is characteristic of relatively deep marine conditions, and possibly appropriate at abyssal depths. However, the constraint offered by the Pressure Compensation Level ((PCL) *c.f.* Fisher 1984; see Chapter 5) associated with contemporaneous volcanism does not support this. For example, the nature of silicic volcanism on Ramsey Island indicates that the marine environment was likely to have been at depths above the Pressure Compensation Level and less than 500m (Kokelaar *et al.* 1985). The depths of extrusion associated with the Fishguard Volcanic Complex are likely to have been both above and below the PCL, possibly in the range 200-1000m {5.4.}. At Abereiddi Bay, hydroclastics appear to dominant over true pyroclastics within the *Murchisoni* Ash and Llanrian Volcanics (Kokelaar *et al.* 1984b) suggesting depths less than 500m. Such constraints indicate that sedimentation during this prolonged period was for the most part above 500m and may have been appreciably less.

If not depth related, the reasons for the development of a contemporaneous mudstone blanket across much of South Wales during late Arenig to Upper Llanvirn times remains largely uncertain. It is increasingly suggested that such conditions represent a restricted basin environment (see Fitches & Woodcock 1987 *Conference report*). Legget (1980) argues that such a model is generally inapplicable to the entire Welsh basin, although he does indicate that such conditions my be appropriate to SW Dyfed from Arenig to Llandeilo times. This is based on the fact that mudstone deposition is not associated with transgression. Therefore there is a tendency to restrict water circulation which would result in density stratification in the deeper parts of the Welsh Basin (i.e. SW Dyfed). From the palaeobathymmetric control offered by contemporaneous volcanism, it is unlikely that the area would have represented a 'deeper' part of the Welsh Basin (if deep parts existed). Depths may well have been within the range of modern shelves. There appears to be as yet, no simple explanation for the development of thick mudstones sequences within SW Dyfed, the entire sequence requiring more detailed work.

4.6. LLANVIRN/LLANDEILO STRATIGRAPHY AND SEDIMENTOLOGY.

Whilst the Arenig-Llanvirn Shales of SW Dyfed remain poorly understood, so to are the sediments believed to be of Llandeilo age. In the following section, strata which may represent the development of the late upper Llanvirn, Llandeilo and basal Caradoc (Constonian Stage) is considered.

4.6.1. Fishguard Bay Group (FBG). *Cumulative thickness (>1500m)*

The Fishguard Bay Group is introduced for sediments believed to range in age from the late Upper Llanvirn to the basal Caradoc. The base of the Group overlies the Fishguard Volcanic Complex at the entrance to Lower Town Harbour (Map 1.). Here arenaceous beds referred to as the Lower Town Formation pass upwards into argillic sediments of the Dyffryn Formation, the latter yielding a fauna indicative of the Lower Llandeilo. Further to the east, around Dinas Head, a thick sequence of predominantly argillic sediments; the 'Llandeilo Shales' of Lowman & Bloxam (1981), here referred to as the '*N. gracilis* Shales', have yielded a fauna indicative of the Upper Llandeilo/Constonian Stage (Lowman 1977). The *N. gracilis* Shales are outside of this study area, although there is little doubt that they represent part of the same thostratigraphic unit. Collectively, the Lower Town Formation, Dyffryn Formation and *N. gracilis* Shales constitute the Fishguard Bay Group, the thickness of which is estimated to be in excess of 1500m (Table 4.1.). The biostratigraphic control offered by the Group, unlike all other sequences, invites a brief discussion on regional correlation.

4 6.1.1. Lower Town Formation (G).*Cumulative thickness (> 220m)*

The Lower Town Formation is seen to crop out on the headlands of Saddle Point [S.M. 9580 3783] and Castle Point [S.M. 9580 3775] at the harbour entrance of Lower Town, Fishguard; from where the Formation can be traced inland for distances exceeding 1km (Map 1.). It is comprised of a thick succession of interbedded coarse sandstones, siltstones, mudstones, and minor volcanoclastics (Fig. 4.6.).

In this study area the Lower Town Formation is fault bounded south of Saddle Point [S.M. 9579 3766], against *D. murchisoni* Shales and rhyolites of the Fishguard Volcanic Complex (Map 1.). The Formation's relationship to the overlying marine mudstones of the Dyffryn Formation is obscured by alluvial cover, although it is realistic to assume that it represents part of a continuous sequence (Fig. 4.6.). The thickness is believed to be in excess of 220m, although extensive strike faulting and discontinuous exposure makes a true estimate indeterminable. On the basis of lithology, the Formation has been divided into two members, the Castle Point and Pantycelyn Members (Table 4.1., Fig. 4.6.). The Castle Point Member relates directly to massive/thick bedded coarse sandstones described by Lowman (1977, his

Castle Point beds), the Pantycelyn Member represents an alternating sequence of mudstones, siltstones, sandstones, volcanoclastics, and conglomerates.

Age constraints on the Lower Town Formation

From beds referred to here as the Castle Point Member, Reed (1895) collected faunal fragments, including *Trinucleus concentricus* var. *favus* (now referred to as *Marrolithus favus*). *M. favus* is a species characteristic of the Upper Llandeilo *Marrolithus favus* Biozone, which according to the scheme of Wilcox & Lockley (1981) is broadly equivalent to the *N. gracilis* Biozone (Table 4.4); tending to suggest that an Upper Llandeilo age is appropriate. A search at the Sedgwick Museum, Cambridge, for the fragments which Reed (1895) collected from Castle Point [S.M. 9579 3766], unfortunately failed to find any ledgered specimens (*writ. comm.* M. Dorling 1986).

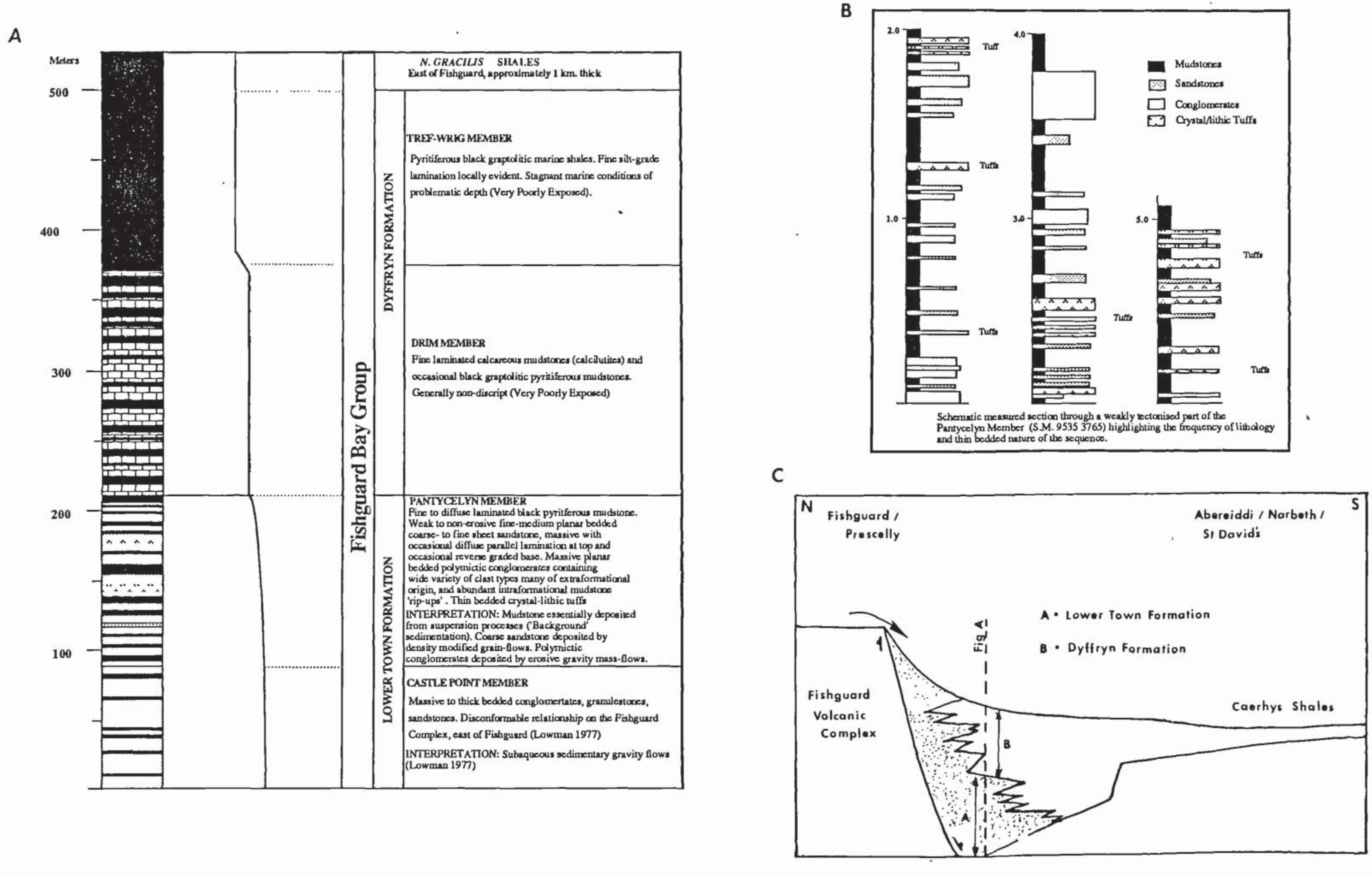
Whilst noting Reed's identification (or possible misidentification), the Lower Town Formation disconformably overlies the Fishguard Volcanic Complex and *D. murchisoni* Shales east of Fishguard (Lowman 1977). It is succeeded by thick marine shales of the Dyffryn Formation which have yielded a graptolite fauna indicative to the Lower Llandeilo *G. retiusculus* Biozone {4.6.2}. Whilst debate and uncertainty still surrounds the exact range of the *murchisoni* and *teretiusculus* biozones{2.3.} this fauna effectively confines the Formation to either the uppermost Llanvirn or lowermost Llandeilo, although of course it may span the Upper Llanvirn/Lower Llandeilo boundary.

a). Castle Point Member (G1). *Type Locality [S.M. 9610 3780, east of this study area];*

The Castle Point Member is introduced for coarse sandstones and subordinate mudstones, relating directly to the *'Castle Point Beds'* as described by Lowman (1977) from Castle Point [S.M. 9610 3780], Lower Town, Fishguard. This locality, whilst outside of this area, is taken as the appropriate type-section due to its accessibility and good exposure. Lowman (1977) defines the Member as a succession of well bedded greywackes, mudstones, conglomerates and breccias, deposited from subaqueous sedimentary gravity-flows (Fig. 4.6.), which rests with angular disconformity on the Fishguard Volcanic Complex, east of Fishguard.

Within this study area the Castle Point Member is poorly exposed, restricted to deeply weathered outcrop on the headland of Saddle Point [S.M. 9570 3770]. Here, the typical lithology can best be described as a series of thick to massive bedded (40cm-1m), compositionally and texturally immature coarse sandstones with minor shale partings. They exhibit features such as weak grading, erosive bases, and abundant intraformational shale clasts. Such features may support a mass-flow origin, however, the sequence has not been studied with regards its sedimentology and the findings of Lowman (1977) are accepted. At

Fig. 4.6. Schematic sedimentary logs and interpretative diagram of the regional stratigraphy. **A.** Composite schematic log of the Fishguard Bay Group. **B.** Frequency diagram of the alternating lithologies within the Pantycelyn Member. **C.** Interpretative schematic section from Fishguard to Abereiddi Bay which may account for variation in lithology, thickness, and stratigraphic relationships.



A
Meters
500
400
300
200
100
Fishguard Bay Group
N. GRACILIS SHALES
East of Fishguard, approximately 1 km. thick
DYFFRYN FORMATION
TREF-WRIG MEMBER
Pyritiferous black graptolitic marine shales. Fine silt-grade lamination locally evident. Stagnant marine conditions of problematic depth (Very Poorly Exposed).
DRIM MEMBER
Fine laminated calcareous mudstones (calcilutites) and occasional black graptolitic pyritiferous mudstones. Generally non-discript (Very Poorly Exposed)
LOWER TOWN FORMATION
PANTYCELYN MEMBER
Fine to diffuse laminated black pyritiferous mudstone. Weak to non-erosive fine-medium planar bedded coarse- to fine sheet sandstone, massive with occasional diffuse parallel lamination at top and occasional reverse graded base. Massive planar bedded polymictic conglomerates containing wide variety of clast types many of extraformational origin, and abundant intraformational mudstone 'rip-ups'. Thin bedded crystal-lithic tuffs
INTERPRETATION: Mudstone essentially deposited from suspension processes ('Background' sedimentation). Coarse sandstone deposited by density modified grain-flows. Polymictic conglomerates deposited by erosive gravity mass-flows.
CASTLE POINT MEMBER
Massive to thick bedded conglomertates, granulestones, sandstones. Disconformable relationship on the Fishguard Complex, east of Fishguard (Lowman 1977)
INTERPRETATION: Subaqueous sedimentary gravity flows (Lowman 1977)
B
2.0
1.0
4.0
3.0
5.0
Tuff
Tuffs
Mudstones
Sandstones
Conglomerates
Crystal/lithic Tuffs
Schematic measured section through a weakly tectonised part of the Pantycelyn Member (S.M. 9535 3765) highlighting the frequency of lithology and thin bedded nature of the sequence.
C
N
S
Fishguard / Prescelly
Abereiddi / Norbeth / St David's
Fig A
A = Lower Town Formation
B = Dyffryn Formation
Fishguard Volcanic Complex
Caerhys Shales
A
B

Saddle Point the Member is fault bounded against *D. murchisoni* Shales and rhyolites of the Fishguard Volcanic Group, although there appears little doubt that it succeeds this sequence.

(B) Pantycelyn Member (G2). *Type Locality (S.M. 9535 3772); Thickness (>80 m).*

The Pantycelyn Member overlies the massive bedded sandstones of the Castle Point Member, although can be regarded as a fine-grained continuum, where mudstone:sandstone ratios are much higher and scale of bedding finer. Typical lithologies consist of alternating thin to medium bedded sheet sandstones, crystal-lithic tuffs and unstructured polymictic conglomerates, interbedded with black pyritiferous mudstones and siltstones. The Member is continuously exposed in a cliff-section below Pantycelyn [S.M. 9546 3774] to the main road [S.M. 9505 3760]. Its heterogeneous character is however, best demonstrated to the east of this section; where numerous gullies and a small wave-cut platform allow the various lithologies to been seen perpendicular to strike. The Member can be traced discontinuously inland for a distance of over a 1km (Map 1.), thinning westwards as a result of faulting against the Fishguard Volcanic Complex. North of Drim Wood [S.M. 9470 3740], the Member is succeeded by laminated calcareous mudstones of the overlying Dyffryn Formation (see below).

Description: The Pantycelyn Member exhibits a wide variety of interbedded lithologies, which can best be considered under the following: laminated mudstone and siltstones, planar bedded coarse sandstones, polymictic conglomerate, and volcaniclastics (Fig. 4.6.). The mudstone-siltstone facies reflects the background sedimentation onto which other lithologies were superimposed, both as individual and locally stacked sequences. Tectonic modification often makes interpretation difficult, although an understanding of the sedimentary history can be gained from both field and petrographic observation.

The *laminated mudstones and siltstones* are in general character to the Arenig-Llanvirn Shales {4.6} as they form a monotonous sequence, although structures which can be interpreted in terms of turbidity current origin are generally absent; or if originally present, obscured by tectonism. The mudstones are highly carbonaceous and pyritiferous, locally containing a faint lamination giving a 'couplet-like' appearance of alternating black carbonaceous mudstone and dark-grey silty shale. A variant of this fine grained 'background' facies, is the local occurrence of subangular to subrounded volcanic boulders 'suspended' as isolated intraclasts in the homogenous matrix (see below).

Planar bedded coarse sandstones occur as discrete buff-white horizons which show lateral continuity through outcrop (where tectonism is minimal). Individual beds are characterised by their textural maturity, although this may be more apparent than real (see below). Bed thickness is variable (4-10cm), and bedding contacts generally planar, although lower and upper surfaces are inevitably tectonically modified; evidence for which is seen by quartz veining and prevalent stylolites defined by irregular trails of unresolved opaque and clay detritus. The sandstones are generally medium to coarse grained and massively bedded. Grading is generally absent, although slight coarse-tail grading above the base of individual beds can be observed in the field, whilst it is common in thin section. Lower surfaces are seen to be weakly scoured in thin section, a feature not readily apparent at outcrop. Diffuse parallel lamination is observed at the top of some units. Some upper surfaces possess a weak asymmetric rippled geometry, although it is uncertain whether this represents tectonic modification, or the result of current modification. Intraclastic mudstones and siltstone rip-ups are a common feature, ranging from small flakes and chips to sub-rounded pebbles. Intraclasts rarely show preference for vertical partitioning in the sandstone bed, although they may show slight concentration above the base. No slumping, dewatering or biogenic structures have been observed.

In thin section the sandstones are medium- to coarse grained, quartz-feldspar dominated, with subordinate lithic- and volcanic rock fragments; better termed subarkose or sublitharenite. Fine grained matrix generally constitutes less than 5% of the rock, although epitaxial quartz overgrowths, reflecting pressure solution accompanying deformation and stylolitisation, make a true estimate of the original matrix and rock composition problematic. However, original grain boundaries are clearly visible by a thin dense pellicle of unresolved opaques, from which a primary rounded to well-rounded morphology is evident, indicating that the sandstones have a shallow marine history. Due to effects of low-grade metamorphism feldspar compositions are now exclusively albite (from microprobe analysis (100% Ab)). Heavy mineral phases are present as accessory components, the most commonly observed being zircon and tourmaline. Tourmaline commonly occurs as subhedral to anhedral grains, generally less than 1mm in diameter, and interstitial with regards the quartz-feldspar framework. Analysis indicate that the composition is intermediate in character between the dravite and schroll end-members, suggesting a metasedimentary origin may be appropriate, rather than the commonly assumed igneous/pegmatitic source (Fig. 4.7.).

Polymictic paraconglomerates occur as discrete planar bedded units. Individual beds vary in thickness from 0.5 – 1m and show laterally continuity through outcrop; although tectonism invariably leads to some degree of discontinuity. Lower and upper bedding contacts are invariably sharp to weakly undulatory. Clast types of variable composition are seen throughout, and include lithic arenites, acidic and more rarely basaltic lava fragments, and intraformational mudstones and laminated siltstones; volumetrically the latter predominate (Plate 4.15.). The lava fragments are commonly of low sphericity, rounded to well rounded pebble to cobbles, strongly suggesting prolonged retention in a shallow marine environment. The clasts were probably supplied from reworking of parts of the underlying Fishguard

Volcanic Complex (see below). Mudstone clasts are generally subangular to sub-rounded, commonly possessing a bladed appearance, much of which is due to tectonic modification, Ragged terminations suggest however, that their origin is one of intra-formational 'rip-ups', an inference supported by the sub-rounded nature of the other component clast types. The matrix is poor to unsorted siltstone and fine sandstone, and constitutes up to 40% of the rock type. A weak bedding planar clast fabric is locally evident, although whether or not this reflects imbrication or subsequent tectonic modification is uncertain.

Fissile *crystal-lithic tuffs* are common throughout the Pantycelyn Member, readily recognised by their bleached grey iridescent colour (Plate 4.15.). They are generally structureless and variable in thickness from 20–70cm. Grading is generally absent, although faint parallel mudstone lamination is evident at the top of some horizons. Mudchips and flakes are frequent, commonly aligned parallel to bedding. Rare fine pumice shards are evident in thin section, whilst idiomorphic feldspar crystals (albite; pure end-member, microprobe analysis 100% Ab) are seen throughout the deposits. The source of the tuffs remains largely uncertain, although similar successions are documented at comparable stratigraphic levels throughout much of SW Dyfed (see below). It is tempting to suggest that they represent part of the same event, or events.

Interpretation: The *laminated mudstones and siltstones* are thought to represent deposition from suspension. Lamination may reflect subtle changes either in climatic factors or fluctuations in the settling rate of fine grained suspended detritus. Irrespective of the mechanism, the mudstones suggest 'stagnant' anoxic marine conditions, outside of the influence of shallow marine processes.

The *planar bedded sandstones* show features common to coarse grained mass-flow deposits, including weakly erosional contacts, subtle grain gradation, and suspended intraformational mudstone clast. Internal evidence for traction structures, such as current ripple lamination are lacking, although high-grain concentration may restrict grain movement resulting in rapid deposition, rather than deposition from wanning-currents (Middleton 1970). The reverse graded base, thin bedding, and medium- to coarse grain size are features characteristic of grain flow deposit (Lowe 1982), although the original textural maturity is not known due to subsequent dissolution. Such deposits require steep slope (18°-25°) rare in subaqueous environments (Vesser 1983). It is likely however, that the sandstones represent deposits which can be broadly umbrelled under the term 'density-modified grain flows' which share features characteristics of grain, debris and turbidity current flows (Lowe 1982). The sandstone source, or the direction of transportation remains largely uncertain, although there is little doubt that they represent shallow marine sandstones redeposited in deeper water .

In character with the sandstones, the *polymictic conglomerates* are interpreted as mass-gravity flows, although the mechanism responsible for deposition is not readily discriminated due to tectonic modification. However, the flows were strongly erosive as indicated by the abundance of intra-formational mudstone clasts incorporated from the sea-floor, suggesting that they were probably high-velocity and high-density currents. The flows were a significant transporting agent of extraformational lithologies, in particular silicic and basaltic volcanics. The origin of the flows cannot be specifically determined, although it appears highly likely, given the nature of the underlying Fishguard Volcanic Complex that they are derived through the reworking of part of the Complex. This may reflect shallowing of marine conditions, during or subsequent too, the development of the Complex. The presence of isolated boulders in mudstone may reflect 'run-off' blocks from such deposits (Prior 1983).

The *crystal-lithic tuffs* are interpreted as single depositional units superimposed on the background sedimentation. The presence of fine mudstone 'rip-ups' suggest they are not 'air-fall' (Plinian) units rather they represent gravity flows, although the presence of fine pumice shards indicate little or no retention in a shallow marine environment.

Depositional environment of the Lower Town Formation: It would appear that the Lower Town Formation represents a complex depositional system; anomalous in comparison to deposits in the immediate vicinity at comparable stratigraphic levels (see below). The background sedimentation of marine mudstones reflects a relatively deep marine environment, outside of the influence of shallow marine processes. However, the detrital mineralogy and grain morphology of the sandstones and the sphericity of cobbles in the conglomerates indicate retention in a shallow marine environment. This disparity is most easily explained in terms of resedimentation and transportation of the clastics by sedimentary gravity flows into it deeper water.

With respect to the depositional system, the deposits may be analogous to a small submarine fault slope-apron facies (Stow 1984), developed as a clastic wedge from uplifted topographic highs possibly situated to the north or east of Fishguard (Fig. 4.6.), rather than to the south (see below). Whilst speculative, such an interpretation accounts for: a) the lithological composition of the deposits, b) the anomalous thickness (i.e. trough-fill at the base of a slope) of the Formation in comparison to surrounding successions to the south, c) the increase in the mudstone-sandstone ratio from the Castle Point Member to the Pantycelyn Member, which may reflect a slope parallel arrangement of coarse to fine facies, d) the presence of clast types which indicate reworking of the Fishguard Volcanic Complex and other strata in a shallow marine environment, e) the overall vertical facies change from the Lower Town Formation into the overlying Dyffryn Formation (see Fig. 4.6.).

Plate 4.15 Polymictic paraconglomerates from the Pantycelyn Member of the Lower Formation [S.M. 9546 3774]. Note abundant intraformational mudstone clasts. A= crystal-lithic tuff. Beds are vertically dipping. Scale is given by a measuring tape which is 17cm in diameter

Plate. 4.16. Typical unweathered calcareous mudstones of the Drim Member, exposed in a temporary drainage cutting along the railway north of Troed-Y-Rhiw [S.M. 9392 3720]. The dark carbonaceous bed in center view, yielded a mixed graptolite fauna indicative of the Lower Llandeilo.

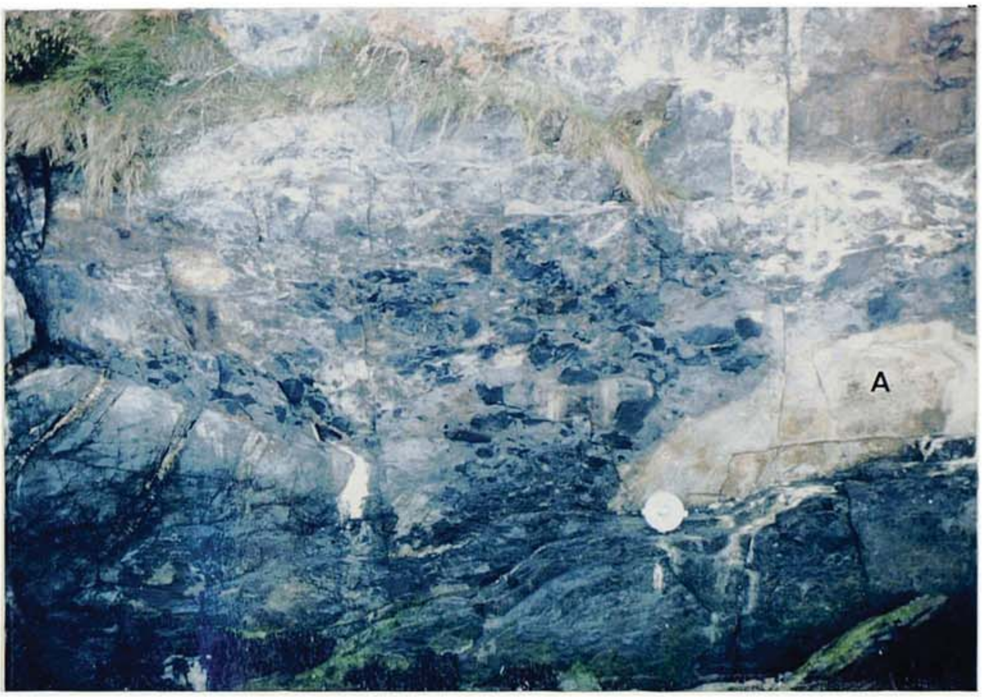
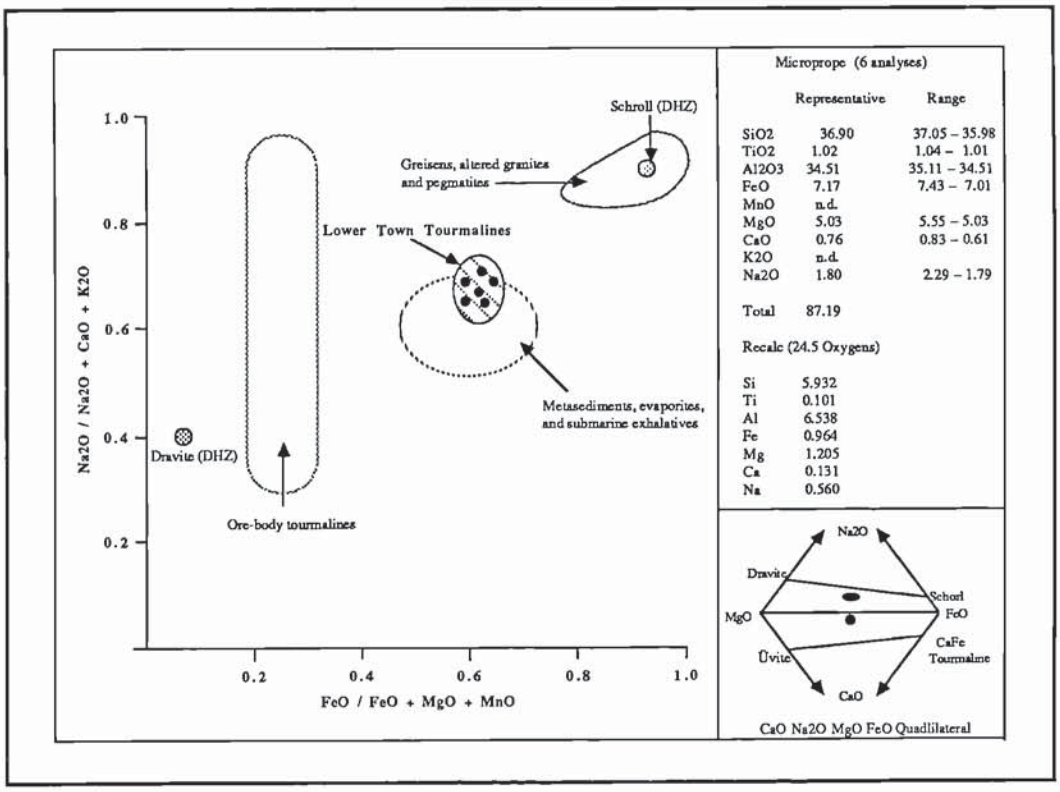
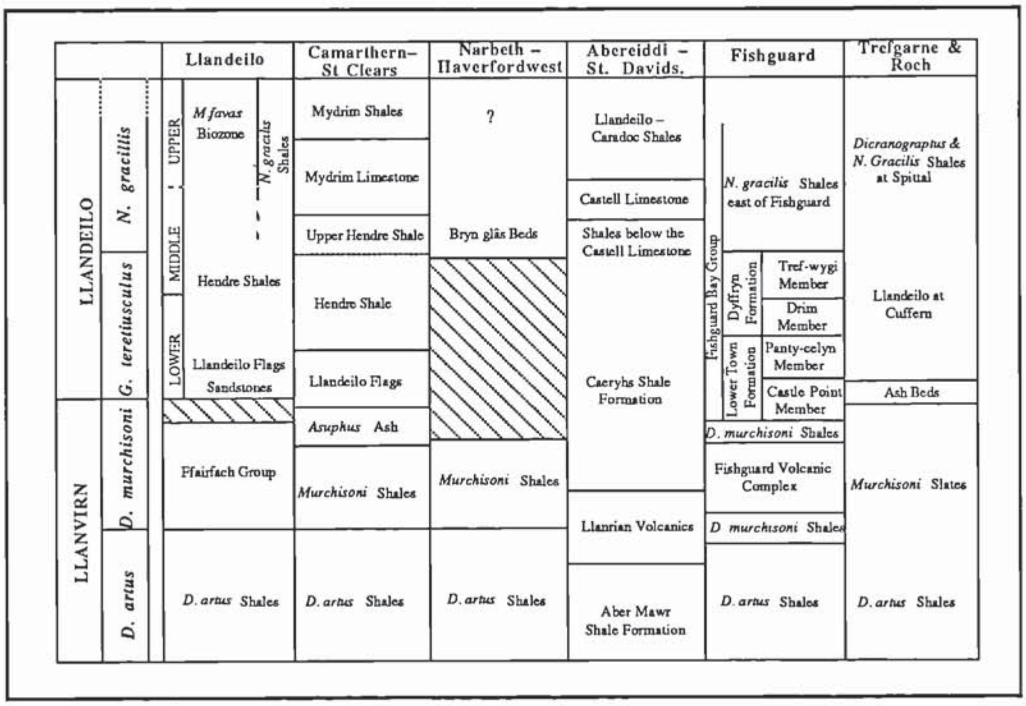

A

Figure 4.7. Comparative Na_2O / (Na_2O + CaO + K_2O) *vs.* FeO / (FeO + MgO + MnO) diagram of detrital tourmaline composition from sandstones within the Pantycelyn Member. Recalculation is based on 24.5 oxygens (anhydrous) with an assumed structural formula $WX_3Y_6(BO_3)_3Si_6$ O18 (OH, F)$_4$, where W= Na, Ca, and K, X= Al, Fe^{2+}, Fe^{3+}, Li, Mg, and Mn, Y= Al, Fe^{3+}; FeO is assumed to be Fe^{2+} (*n.b.* B_2O_3, F, Li_2O are indeterminate in part accounting for Wt.% oxide totals less than 100). Compositional fields of igneous tourmalines after Plimmer (1983 and references therein) sedimentary and ore-body tourmaline after Either & Campbell (1977), Plimmer (1983), Uitterdijk Appel (1984). Dravite and Schrol end-members compositions (DHZ) are from Deer *et al.* (1962).

Table. 4.4. Llandeilo and Upper Llanvirn stratigraphy in SW Dyfed.


Na2O / Na2O + CaO + K2O
FeO / FeO + MgO + MnO
1.0
0.8
0.6
0.4
0.2
0.2
0.4
0.6
0.8
1.0
Schroll (DHZ)
Greisens, altered granites and pegmatites
Lower Town Tourmalines
Metasediments, evaporites, and submarine exhalatives
Dravite (DHZ)
Ore-body tourmalines
Microprope (6 analyses)
Representative
Range
SiO2 36.90 37.05 – 35.98
TiO2 1.02 1.04 – 1.01
Al2O3 34.51 35.11 – 34.51
FeO 7.17 7.43 – 7.01
MnO n.d.
MgO 5.03 5.55 – 5.03
CaO 0.76 0.83 – 0.61
K2O n.d.
Na2O 1.80 2.29 – 1.79
Total 87.19
Recalc (24.5 Oxygens)
Si 5.932
Ti 0.101
Al 6.538
Fe 0.964
Mg 1.205
Ca 0.131
Na 0.560
Na2O
Dravite
Schorl
MgO
FeO
Uvite
CaFe Tourmaline
CaO
CaO Na2O MgO FeO Quadlilateral


Llandeilo
Camarthern– St Clears
Narbeth – Haverfordwest
Abereiddi – St. Davids.
Fishguard
Trefgarne & Roch
LLANDEILO
LLANVIRN
N. gracilis
G. teretiusculus
D. murchisoni
D. artus
UPPER
MIDDLE
LOWER
M favas Biozone
N. gracilis Shales
Hendre Shales
Llandeilo Flags Sandstones
Pfairfach Group
D. artus Shales
Mydrim Shales
Mydrim Limestone
Upper Hendre Shale
Hendre Shale
Llandeilo Flags
Asaphus Ash
Murchisoni Shales
D. artus Shales
?
Bryn glâs Beds
Murchisoni Shales
D. artus Shales
Llandeilo – Caradoc Shales
Castell Limestone
Shales below the Castell Limestone
Caeryhs Shale Formation
Llanrian Volcanics
Aber Mawr Shale Formation
N. gracilis Shales east of Fishguard
Fishguard Bay Group
Dyffryn Formation
Tref-wygi Member
Drim Member
Lower Town Formation
Panty-celyn Member
Castle Point Member
D. murchisoni Shales
Fishguard Volcanic Complex
D murchisoni Shales
D. artus Shales
Dicranograptus & N. Gracilis Shales at Spittal
Llandeilo at Cuffern
Ash Beds
Murchisoni Slates
D. artus Shales

Correlation of the Lower Town Formation

As outlined above, the Lower Town Formation is likely to be of uppermost Llanvirn or lowermost Llandeilo in age, or spans the boundary. The Llanvirn–Llandeilo boundary is poorly understood lithostratigraphically and debate still surrounds the exact biostratigraphic range of the *murchisoni* and *teretiusculus* biozones {2.3.}. Nevertheless, the Formation represents a change in the nature of sedimentation and invites comparison with successions from Llandeilo, Carmarthen, Abereiddi Bay and Narbeth districts, where strata of comparable stratigraphic age are recorded.

Abereiddi Bay and Narbeth districts: To the south-west of Fishguard, little resembling the Lower Town Formation is evident. At Abereiddi Bay, black graptolitic shales of the Caerhys Shale Formation pass upwards from the *D. murchisoni* Ash at the base of the *murchisoni* Biozone, to the Castell Limestone of the *gracilis* Biozone (Hughes *et al.* 1982). The Upper Llanvirn – Llandeilo transition is recognised entirely on faunal grounds; there being no evidence for a change in the style of sedimentation or an influx of coarse clastic sediments (see la er). Further to the south around Narbeth, it has been suggested that shelly Llandeilo lımestones and flags of the *gracilis* Biozone rest on shales of the *artus* Biozone, both the *murchisoni* and *teretiusculus* biozones being absent (Addison 1974). Recently, marine black shales with a mixed graptolite-trilobite fauna indicative of the low *murchisoni* Biozone (Lan M ll Shales, Table 4.4.) have been identified around Narbeth, underlying the Bryn-glâs Beds (Kennedy 1986). However, sediments of Lower Llandeilo age appear to be absent in this area. Thıs may suggest a thinning of the Upper Llanvirn/Lower Llandeilo southwards, in apparent contradiction to the 'classic' Welsh Basin model. Speculatively, a small graben system may alleviate some of this disparity and accommodate the present understanding of the stratigraphy and the sedimentology of the Llanvirn/Lower Llandeilo in this part of SW Dyfed (Fig. 4.6.c.).

Carmarthen and Llandeilo districts: In the stratotype districts of Llandeilo-Llangadog east of Carmarthen, the Llandeilo Series is underlain by the Ffairfach Group. This consists of shelly shallow marine arenaceous-volcanoclastic facies which had been assigned to the base of the Llandeilo (Strathern *et al.* 1909), although subsequently shown to be of Upper Llanvirn age (Williams 1953). The Group is approximately 180m thick, and comprises a heterolithic sequence of conglomerates, grits, sandstones, flags and crystal tuffs, which pass upwards with apparent conformity into the Llandeilo (Williams 1953, 1969; Wilcox & Lockley 1981, Bassett 1982). It has generally been accepted that the Ffairfach Group is a rather anomalous shallow marine facies within the *murchisoni* Biozone, as mudstone deposition dominated the style of sedimentation throughout much of the South Wales during the Upper Llanvirn.

However, west of Carmarthen around the Mydrim district, Catterall & Thomas (in: Strathern *et al.* 1909), correlate pyroclastic and arenaceous deposits of the *Asaphus* Ash, with ashes in the upper most parts of the Ffairfach Group. Both the *Asaphus* Ash and the Ffairfach Group are overlain by siltstones and mudstones collectively referred to as the Llandeilo Flags which contain a faunal assemblage diagnostic of the *teretiusculus* Biozone (Toghill 1970, Williams *et al.* 1972). This volcanogenic influence at both Mydrim and Llandeilo may be represented in the Fishguard district by the crystal-lithic tuffs in the Lower Town Formation.

In would appear therefore, that both the successions at Llandeilo and Mydrim have sufficient similarities in both lithology and associated surrounding strata to invite correlation with the Lower Town Formation (Table 4.4); it being likely that they reflect a part of a distinctive regional change at the top of the Llanvirn and Lower Llandeilo. Wilcox & Lockley (1981) suggest that the transition from the Ffairfach Group to the Llandeilo Flags relates to a lower Llandeilo transgression. However, an alternative explanation for the coarse clastic facies at the top of the Llanvirn/Lower Llandeilo (as seen at Fishguard, Mydrim, Carmarthen) and Llandeilo, is that it represents a temporary change in the regional basin dynamics during the late Llanvirn, unrelated to a transgressive event.

6.4.1.2. The Dyffryn Formation (H). *2 Members; Thickness (?> 550 m).*

Overlying the Lower Town Formation, a thick succession of calcareous and pyritiferous black marine shales referred to collectively as the Dyffryn Formation occupy the area between Fishguard and Goodwick. The Formation relates in part to successions previously referred to by Cox (1930) and Thomas & Thomas (1956) as *'Calcareous flags of Llandeilo and Bala age'*. Faunal evidence suggests that the successions are restricted to the *G. teretiusculus* Biozone indicating a Lower Llandeilo age. Based on varying lithology and fauna the Formation is divided into two members termed here the Drim and Tref-wrgi Members. The following section, whilst primarily concerned with the Dyffryn Formation, also considers the *N. gracilis* Shales east of Fishguard (Lowman & Bloxam 1981), which are the likely to represent the along strike continuum.

Exposure of the Dyffryn Formation is extremely poor, being restricted to shallow grassy exposures on the upstanding knoll of 'The Drim' [S.M. 9445 3720], infrequent drainage and railway cuttings and two quarry exposures. The Formation is fault bounded to the northwest against rhyolites of the Fishguard Volcanic Complex (Map 1.), whilst to the southeast its base is nowhere exposed. It is however, realistic to assume that the Formation lies above the Pantycelyn Member which 'tectonically' thins to the south west, and brings into contact the Volcanic Complex and the Dyffryn Formation in the vicinity of Comins Cwmbrandy [S.M. 9445 3703] and Mill Farm [S.M. 9400 3685]. Poor exposure, and evidence for intense deformation, make a true estimation of the Formation's thickness indeterminable. A thickness of 550m is thought appropriate assuming no fault repetition and a constant northwards younging. The Formation's stratigraphic position, and distribution around Fishguard and Goodwick, suggests that the low-lying marshy ground in the vicinity of Manorowen and further to the south-west around Carn Ilis is also occupied by the Formation.

(a). Drim Member (H1). *Type Locality (S.M. 9329 3729), Thickness (> 350m).*

The Drim Member represents a distinctive (although poorly exposed) succession, of thin bedded calcareous mudstones which overlie the Pantycelyn Member of the Lower Town Formation, north of Fishguard. This Member relates in part to the *'Drim Wood Flags'* of Cox (1930) and *'Calcareous Flags'* of Thomas & Thomas (1956). The Drim Member is best observed at two localities: a small quarry [S.M. 9457 3730] at the eastern edge of 'The Drim', and in intermittent exposure along the railway cutting between Cwm Celyn [S.M. 9403 3750] and Crowstone [S.M. 9385 3718]. Weathering at these localities is invariably deep, and accentuated by the strong penetrative cleavage, surfaces being stained brown to ocherous the result of pyrite oxidation. A locality where relatively fresh surfaces can be observed, occurs b low the railway bridge at the top of Troed-Y-Rhiw [S.M. 9392 3720]. This is taken to be an appropriate lithotype section, although only a few meters are exposed; limited exposure also occurs in the shallow cutting along the railway line further to the west.

The typical unweathered lithology is one of thin bedded (1-5cm) calcareous siltstones and mudstones (calcilutites), interbedded with graptolitic black pyritiferous mudstones which give a 'banded' appearance to the rock type (Plate 4.16.). Sedimentary structures are limited and diffuse. Fine parallel lamination is by far the most common sedimentary structure to the black mudstones, possibly reflecting seasonal variation and deposition from suspension processes. The carbonate beds are generally deeply weathered, offering little in the way of recognisable sedimentary structures. It would appear that the carbonate is for the most part a primary feature as it follows bedding surfaces, although locally it obscures boundaries occurring in patchy domains suggesting partial dissolution, possibly by cleavage development. The preservation of graptolites would tend to indicate the development of stagnate environment well away from wave or current agitation. The presence of carbonates suggests that deposition was above the Carbonate Compensation Depth (CCD), although this has little palaeobathymmetric significance. There is no evidence for mass flow deposition or for a significant terrigenous component such as that seen in the underlying Lower Town Formation.

During the summer of 1985, a shallow temporary drainage cutting was made along the railway line from Troed-Y-Rhiw northwards for approximately 200m, exposing a near continuous section oblique to strike though the Drim Member. The lithology was seen to change little, although importantly, a thin pyritiferous black mudstone horizon (Plate 4.16.), yielded a varied graptolite assemblage containing, *Amplexograptus* s.p., *Glyptograptus euglyphus* (Lapworth) s.l., *Glyptograptus teretiusculus* and *Corynoides* s.p. (*pers. comm.* Dr. R.B. Rickards 1986). This assemblage places the Drim Member in the *G. teretiusculus* Biozone and almost certainly a Lower Llandeilo age (Table 4.4.). The occurrence of a graptolite assemblage indicative of the *teretiusculus* Biozone, is extremely useful in defining the thick shale successions that occupy the core of the Goodwick Syncline. Previously, it having been suggested that the successions were of likely Llandeilo (Bala) age, although the successions had not afforded fauna. However, the thickness of this Lower Llandeilo succession is anomalous in comparison to successions elsewhere in SW Dyfed (see below).

(b). Tref-wrgi Member (H2). *Type Locality [S.M. 9380 3769]; Thickness (>200 m).*

The Tref-wrgi Member represents a succession of graptolitic black marine mudstones exposed at only one locality (*n.b.* access to this quarry should only be made once permission is granted by the owners who reside in the nearby motel): the large disused brick quarry at Goodwick [S.M. 9380 3769] north-east of Tref-wrgi Farm [S.M. 9355 3747]. The Member relates in part to the *'Goodwick Shales'* of Cox (1930) and Thomas & Thomas (1956), who equated this succession to the Upper Llandeilo and basal Caradoc (or Bala).

The Tref-wrgi Member is fault bounded to the north against rhyolites of the Fishguard Volcanic Complex, the disparity between both lithologies being highlighted by the Goodwick Fault Scarp, possibly the most prominent feature in this study area. The relationship with the Drim Member is not seen, although, the Tref-wrgi Member must overly this succession and occupy a considerable part of the low-lying marshy area between Cwm Celyn [S.M. 9385 3755] westwards to Carne-coch [S.M. 9260 3715, Map 1.]. The thickness of the Member exposed at the quarry is in excess of 90m, although its true thickness from the Goodwick Fault Scarp to the top of the Drim Member is likely to be in excess of 200m.

The typical lithology of the Tref-wrgi Member is one of fissile homogenous black marine shales, similar in character to those interbedded with the calcareous mudstones of the Dyffern Member. They commonly show evidence for fine parallel lamination, locally giving the appearance of a rhythmite bedded mudstone. Pyrite is abundant as fine grained disseminated rhombs, locally concentrated along bedding surfaces, and more rarely as large concretions within highly carbonaceous horizons. Graptolitic horizons are locally evident and fauna prolific, although bedding/cleavage relationships are generally unfavorable for the collection of undeformed specimens. However, on the most southerly facing quarried surface at the west end of the quarry [S.M. 9375 3765], abundances of relatively large undeformed graptolites can be obtained. From this locality, the graptolite assemblage is highly restrictive in nature, only *G.*

teretiusculus being identified (*pers. comm.* Dr. R. B. Rickards 1986). Whilst the assemblage is monospecific, the abundance of *G. teretiusculus*, the absence of pendant didymograptids, and a stratigraphic relationship above the Drim Member, almost certainly places the Tref-wrgi Member in the Lower Llandeilo. If correct, then it would appear that the development of Lower Llandeilo sediments in the Fishguard district, are anomalous in their thickness by comparison to other Lower Llandeilo successions in South Wales. The development of the Llandeilo and probably Caradoc throughout the area may have a cumulative thickness in excess of 2km, representing the thickest accumulation of such strata in South Wales.

Interpretation

The Dyffryn Formation is interpreted as representing deposits which accumulated essentially from suspension, in a similar manner to the preceding Arenig-Llanvirn Shales. Evidence for structured depositional units similar to the mud-turbidites of the Penmaen Dewi Formation are absent. Fine parallel lamination, and faint cross-laminated siltstones are the only recognisable sedimentary structures and may represent distal turbidites, although unfavourable bedding/cleavage relationships distort sedimentary structures. The abundance of pyrite and local preservation of graptolites probably reflects anoxic and stagnant marine conditions, although their significance is limited in the absence of biogenic and ordered sedimentary structures.

c). *N. gracilis* Shales (outside of this study area).

To the east of Fishguard overlying the Lower Town Formation, Lowman & Bloxam (1981), refer to a thick succession of shales (which they term the 'Hendre Shales' following Evans 1944), as being of Middle Llandeilo to Constonian (basal Caradoc) age. This age is based on the occurrence of *N. gracilis* within assemblages that they obtained from Aber Howell [S.M. 9903 3850], near Dinas Head. It seems highly probable, given the close proximity of both study areas, that at least part of the thick shale successions south of Aber Howell are easterly equivalents of the Dyffryn Formation. If this is the case, then the Llandeilo successions in the Fishguard – Dinas Head district are likely to have a cumulative thickness in excess of 1.5 km; whilst identification of both the *teretiusculus* and *gracilis* biozones in the Fishguard – Dinas Head districts, probably indicates the total development of the Llandeilo, and may repay further investigation, in that it may prove to be a useful stratotype locality for the Llandeilo in this part of Wales.

Correlation of the Dyffryn Formation and N. gracilis Shales

Cox (1930) was of the opinion that the shales and calcareous mudstones described here as the

Dyffryn Formation, where best equated with the '*Dicranograptus* Shales' of Abereiddi Bay (Cox 1930, Cox *et al.* 1930a), a correlation justified in the absence of diagnostic faunal evidence. The '*Dicranograptus* Shales', recently renamed the 'Llandeilo and Caradoc Shales' (Hughes *et al.* 1982) are regarded however, as being of Upper Llandeilo and lowermost Caradoc in age. Their lowest beds are confined to the *gracilis* Biozone, it therefore likely that they equate with the *N. gracilis* Shales east of Fishguard. The identification here of the Dyffryn Formation being confined to the Lower Llandeilo *teretiusculus* Biozone, invites a reassessment of previously equated Llandeilo successions in SW Dyfed; chiefly from the, Roch–Trefgarne–Abereiddi Bay-Narbeth and Mydrim-Carmarthen-Llandeilo districts (Table 4.4.).

Roch-Trefgarne-Abereiddi Bay–Narbeth: To the south around Trefgarne and Roch, Thomas & Cox (1924) record the development of shales which appear to range from the *gracilis* Biozone to the basal Caradoc (*Dicranograptus* Shales) and are likely to correlate with the *N. gracilis*' shales east of Fishguard. At Cuffern, near Roch, Kennedy (1986) records a mixed trilobite, brachiopod and graptolite fauna of Lower Llandeilo aspect within shales that overlie '*D. murchisoni* Slates' tentatively equating the succession to the lower parts of the Llandeilo (Table 4 .). If correctly assigned, then the Llandeilo developed at Cuffern, Roch, and Trefgarne, appears to be directly related to the Dyffryn Formation. However, the relationship of the Llandeilo successions with those of the *murchisoni* Biozone is poorly understood, whilst the nature of the sediments and their distribution requires further investigation.

It seems likely that shales of the *gracilis* Biozone crop out extensively in the Prescelly Hills, where Evans (1944) recorded the presence of *Trinucleus* cf. *fimbriatus* Murchison, a species generally restricted to the *gracilis* Biozone (Thomas *et al.* 1984). Evans (1944) further described a thick succession of shales (variously termed the Fenni Fawr Beds, Glouge Slates and Mydrim Slates) which probably relate to the Upper Llandeilo and Lower Caradoc, and have a cumulative thickness in excess of 750m. From such evidence, it would appear that Llandeilo and Caradoc successions in this part of SW Dyfed are well represented and anomalous in their thickness (i.e > 2 km). If thicknesses are correct then they represent by far the thickest group of sediments in SW Wales.

Elsewhere, Hughes *et al.* (1982) identified a 20m interval at the top of the Caerhys Shale Formation and below the Castell Limestone on the northerly inverted limb of the Abereiddi Bay Syncline, which they equate with the *teretiusculus* Biozone. This biozonal correlation is essentially based on the absence of pendant didymograptids and the occurrence of a mixed faunal assemblage including: *Geragnostus* sp., *Platycalymene tasgarensis-simulata*, *Ogyiocarella* cf. *angustissima*, *Protolloydolithus* sp., an assemblage which they suggest possess a Lower Llandeilo aspect. However, renewed collecting from the locality described by

Hughes *et al.* (1982) as belonging to the Lower Llandeilo, has yielded a faunal assemblage indicative of the Upper Llandeilo, including: *Platycalymene tasgarensis-simulata*, *Ogyiocarella debuchii*, *Barrandia* sp., *Trinucleus fimbriatus*, *Segmentagostus mcCoyii*, and *Cnemoidopyge* cf. *parva* (R.J. Kennedy, unpublished data). This apparent discrepancy in faunal assemblages is as yet unresolved, although the disparity in the thickness of the *teretiusculus* Biozone (≈20m) of the Caerhys Shale Formation's, and that of the Dyffryn Formation (≈ 550m) which appears to be confined to the *teretiusculus* Biozone, makes correlation difficult. Lithologically, sediments of the Caerhys Shale Formation are similar to those of the Tref-wrgi Member with which they are here tentatively equated (Table 4.4.), although no strata resembling the Drim Member are seen at Abereiddi Bay. It is possible that the strike fault described by Cox (1916) and Waltham (1971, p.51) as occurring to the north of the Castell Limestone (*gracilis* Biozone) on the northerly inverted limb of the Abereiddi syncline, has caused omission of stratigraphy. However, Black *et al.* (1971) and Hughes *et al.* (1982) make no reference to the existance of this structure.

As outlined earlier, sediments of the *G. teretiusculus* Biozone appear to be absent to the south around the Narbeth district. In this area the lower Upper Llanvirn Lan Mill Shales appear to underlie the Bryn-glâs Beds of the *gracilis* Biozone (Kennedy 1986), and there appears to be no obvious stratigraphic parrallel with the Dyffryn Formation directly south of Fishguard, in terms of thickness or nature of sediments. In respect of the under lying Lower Town Formation, successions to the south of Fishguard such as the Caerhys Shales show no evidence for any clastic influence. This is unusual as it would form the obvious source area for such material under the 'classic' Welsh Basin, basin-shelf morphology of the area. This would tend to give credibility to the scinario of shallow marine conditions to the north of the area filling a graben structure to the south (Fig 4.6.).

Mydrim–Carmarthen: In character with the Lower Town Formation, the best successions with which the Dyffryn Formation can be favourably correlated (both lithologically and bio-stratigraphically), crop out to the east around Mydrim west of Carmarthen. The succession here also accords well with the type succession at Llandeilo (see Toghill 1970). The base of the Llandeilo succession at Mydrim (Table 4.4.) comprises 'Llandeilo Flags' which contain *G.teretiusculus* and a faunal assemblage indicative of the *teretiusculus* Biozone. These overlie the *Asaphus* Ash which is assumed to be of Upper Llanvirn in age if correctly correlated with the Ffairfach Group at Llandeilo. The Llandeilo Flags pass upwards into a thick succession of mudstones termed the 'Hendre Shales'. Jones (1956) suggested that they contained *N. gracilis*, although Toghill (1970) disputed such findings, obtaining an abundant graptolite fauna which he regarded as indicative of the *teretiusculus* Biozone. Addison (1974) however, whilst accepting the findings of Toghill (1970), recorded *N. gracilis* from the uppermost parts of the Hendre Shale below the Mydrim Limestone; the Mydrim Limestone having had previously been known to be restricted to the *gracilis* Biozone and correlated with the Castell Limestone at Abereiddi Bay. It seems likely in the light of the faunal and lithological similarity between the Dyffryn Formation and the 'Llandeilo Flags' around Mydrim, and possibly the 'Llandeilo Flags' which overlie the Ffairfach Group at the Llandeilo type locality (Table 4.4.), that the successions represent a significant and continuous stratigraphic unit within the Lower Llandeilo in this part of S Dyfed.

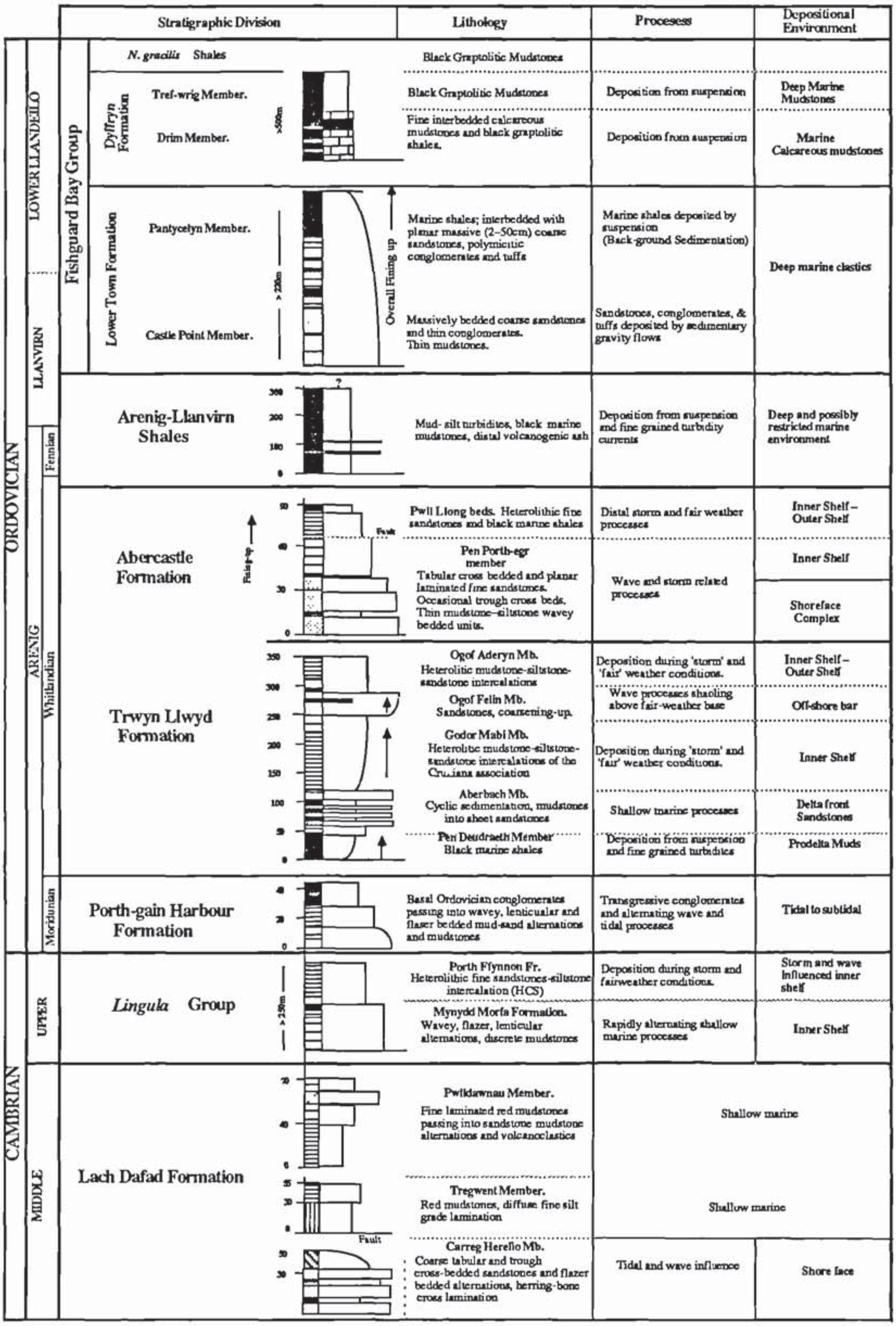
However, equated successions (i.e. upper parts of the Caerhys Shales, Bryn-glâs beds) to the south of these districts appear to be dramatically thinned. This is speculatively interpreted to reflect original basin morphology, although not in the classic sense of the Welsh Basin, in so much as the area to the north formed a topographic high with successions to the south representing accumulation on a shallow 'stable' platform (Fig. 4.6.c.). Such an interpretation is the simplest explanation for thickness variation from north to south, and also takes into account the Lower Town Formation and unroofing of the Fishguard Volcanic Complex.

4 7. SUMMARY

The sedimentary sequence exposed throughout the Fishguard – Porth-gain district is thought to range in age from the Middle Cambrian (St. Davids Series) to the Lower Llandeilo (*G. teretiusculus* Biozone). Directly to the east of Fishguard, the Upper Llandeilo/Caradoc is developed (*N. gracilis* Biozone). It has been found possible to differentiate the succession into 7 principle lithostratigraphic units, within which 23 subdivisions are presented, variously at Formational and Member levels (Table 4.1.). Figure 4.8. is a schematic sedimentary log which summarises many of the findings presented. Correlation with Lower Palaeozoic successions elsewhere in SW Dyfed is given within the text and is summarised in Table (7.1.), which is pre ented at the end of this thesis. Aspects of the regional palaeogeography have been addressed in passing, although few conclusions can be drawn, in so much as, regional correlation is not sufficiently sensitive, and many of the equated successions are poorly understood sedimentologically. In proposed stratigraphic order, the principle lithostratigraphic divisions can be briefly summarised as follows:

Lach Dafad Formation; The lowermost Cambrian (? Middle Cambrian) succession within the area is defined by the Lach Dafad Formation. The Formation is thought to best equate with the lower Middle Cambrian Solva Group developed around the St. Davids district; sediments of the Lower Cambrian Caerfai Group appears to be absent. The Formation defines, tectonically fragmented red-green sandstone-shale successions and associated strata. The lithological

Figure 4.8. Composite schematic sedimentary log of the sedimentary sequence developed throughout the Fishguard – Porth-gain district.



Stratigraphic Division
Lithology
Processess
Depositional Environment
ORDOVICIAN
LOWER LLANDEILO
LLANVIRN
ARENIG
Fennian
Whitlandian
Moridunian
Fishguard Bay Group
N. gracilis Shales
Black Graptolitic Mudstones
Dyffryn Formation
Tref-wrig Member.
>500m
Black Graptolitic Mudstones
Deposition from suspension
Deep Marine Mudstones
Drim Member.
Fine interbedded calcareous mudstones and black graptolitic shales.
Deposition from suspension
Marine Calcareous mudstones
Lower Town Formation
Pantycelyn Member.
> 220m
Overall Fining up
Marine shales; interbedded with planar massive (2–50cm) coarse sandstones, polymictic conglomerates and tuffs
Marine shales deposited by suspension (Back-ground Sedimentation)
Deep marine clastics
Castle Point Member.
Massively bedded coarse sandstones and thin conglomerates. Thin mudstones.
Sandstones, conglomerates, & tuffs deposited by sedimentary gravity flows
Arenig-Llanvirn Shales
300
200
100
0
?
Mud- silt turbidites, black marine mudstones, distal volcanogenic ash
Deposition from suspension and fine grained turbidity currents
Deep and possibly restricted marine environment
Abercastle Formation
Fining-up
50
Fault
Pwll Llong beds. Heterolithic fine sandstones and black marine shales
Distal storm and fair weather processes
Inner Shelf – Outer Shelf
Pen Porth-egr member
Tabular cross bedded and planar laminated fine sandstones. Occasional trough cross beds. Thin mudstone–siltstone wavey bedded units.
Wave and storm related processes
Inner Shelf
Shoreface Complex
30
0
Trwyn Llwyd Formation
350
300
250
200
150
100
50
0
Ogof Aderyn Mb. Heterolitic mudstone-siltstone-sandstone intercalations
Deposition during 'storm' and 'fair' weather conditions.
Inner Shelf – Outer Shelf
Ogof Felin Mb. Sandstones, coarsening-up.
Wave processes shaoling above fair-weather base
Off-shore bar
Godor Mabl Mb. Heterolitic mudstone-siltstone-sandstone intercalations of the Cruziana association
Deposition during 'storm' and 'fair' weather conditions.
Inner Shelf
Aberbach Mb. Cyclic sedimentation, mudstones into sheet sandstones
Shallow marine processes
Delta front Sandstones
Pen Deudraeth Member Black marine shales
Deposition from suspension and fine grained turbidites
Prodelta Muds
Porth-gain Harbour Formation
0
Basal Ordovician conglomerates passing into wavey, lenticualar and flaser bedded mud-sand alternations and mudstones
Transgressive conglomerates and alternating wave and tidal processes
Tidal to subtidal
CAMBRIAN
UPPER
MIDDLE
Lingula Group
> 250m
Porth Ffynnon Fr. Heterolithic fine sandstones-siltstone intercalation (HCS)
Deposition during storm and fairweather conditions.
Storm and wave influenced inner shelf
Mynydd Morfa Formation. Wavey, flazer, lenticular alternations, discrete mudstones
Rapidly alternating shallow marine processes
Inner Shelf
Lach Dafad Formation
70
40
Pwlldawnau Member. Fine laminated red mudstones passing into sandstone mudstone alternations and volcanoclastics
Shallow marine
35
30
Tregwent Member. Red mudstones, diffuse fine silt grade lamination
Shallow marine
Fault
50
30
Carreg Herefio Mb. Coarse tabular and trough cross-bedded sandstones and flazer bedded alternations, herring-bone cross lamination
Tidal and wave influence
Shore face

character of the Formation is pieced together from three fault bounded sequences referred to as the Carreg Herefio Member, Tregwent Member, and Pwlldawnau Member (Table 4.1.). All members appear to be shallow marine in character. The Carreg Herefio Member is of interest, in so much as, deposition was under tidal influence. If the age constraint is correct (to the point that the Formation is of Lower or Middle Cambrian age) this would indicate shallow marine/shoreface conditions existed significantly further north than the present understanding of the regional Cambrian palaeogeography would tend to allow under the 'classic' Welsh Basin model. The Pwlldawnau Member represent an enigmatic sequence, as it contains a volcanoclastic mass-flow which would suggests 'subaerial' volcanism in this part of SW Dyfed during early Cambrian times.

Lingula Group: The Upper Cambrian is represented in the Fishguard – Porth-gain district by sediments collectively referred to as the *Lingula* Group. The Group may be restricted to the Maentwrog and possibly Festiniog Stages of the Merioneth Series; higher beds of the Dolgelly Stage appear to be absent. The intervening Menevian Group which separates the lower Middle Cambrian from the Upper Cambrian to the south of this area, appear not to be represented. The *Lingula* Group is divided into the Porth Ffynnon and Mynydd Morfa Formations. Both Formations are shallow marine in character. The Porth Ffynnon Formation showing features typical of storm and/or wave influenced inner shelf sedimentation, the Mynydd Morfa Formation appears to reflect a wave dominated shelf environment. The entire *Lingula* Group r flects sedimentation on a stable and possibly extensive siliciclastic shelf.

Tremadoc: Sediments of Tremadoc age appear to be absent from the Fishguard – Porth-gain district. The closest documented Tremadoc strata, crop out around Carmarthen (Cope *et al.* 1979). The absence of the development of the Tremadoc has been attributed to uplift during late Tremadoc times with Ordovician marine sedimentation commencing during early Arenig times (Fortey & Owens 1987). However, the exact extent of Arenig erosion is uncertain, and in the apparent absence of Festiniog and Dolgelly beds, it is reasonable to suggest that the non-sequence in SW Dyfed is of far greater duration (i.e. mid-Upper Cambrian to Lower Arenig). Subsequent Arenig sedimentation appears more complex than simply 'flooding' a pre-existing Tremadoc landmass.

Porth-gain Harbour Formation: The onset of Ordovician sedimentation in the Fishguard – Porth-gain district is recorded by sediments of the Porth-gain Harbour Formation, the age of which is argued to be of the lower Arenig Moridunian Stage. The Formation rests with significant hiatus upon the Porth Ffynnon Formation (i.e. Upper Cambrian), and is typically transgressive in character recording deposition following late Upper Cambrian – Tremadoc uplift. The Formation progressively fines up from shallow marine polymictic conglomerates and heterolithic sandstone-mudstone intercalations (Sandstone member) into sublittoral mudstones (Mudstone member).

Abercastle Formation: The Abercastle Formation is known to be of Whitlandian age (Fortey & Owens 1987). From the fragmented successions it is possible to obtain a broad appreciation of the Formation's sedimentary history. The Formation is divided into the Pen Porth-egr Member and Pwll Llong beds. The Pen Porth-egr member is dominated by shoreface sandstones which progressively pass upwards into wave dominated inner shelf siliciclastic. The Pwll Llong beds reflect the transition from siliciclastic sedimentation into mudstones of the Penmaen Dewi Formation. Collectively the Formation reflects a steady deepening marine sequence of shoreface – shallow shelf clastics – shelf mudstones.

Trwyn Llwyd Formation: The Trwyn Llwyd Formation is thought to be of Whitlandian age. The Formation is pieced together from tectonically fragmented repeat sequences. The Formation has a complex storm and wave influnced siliciclastic history. The Formation is divided into 5 members, transitions between which are gradational. The lowermost Member, the Pen Deudraeth Member, is mudstone dominated and interpreted to represent a prodelta/shelf mudstone succession. The Aberbach Member is characterised by coarsening-up cycles in erpreted as deposition with a delta-front sandstone facies. The Godor Mabli Member reflects deposition within a storm and/or wave dominated inner shelf. The Ogof Felin Member is interpreted to reflect a shoaling sandstone facies analogous to an off-shore bar complex. The overlying Ogof Aderyn Member reflects a return to shallow marine shelf sedimentation. The Ogof Aderyn Member passes into the mudstones of the Penmaen Dewi Formation, the transition between which reflects a major change in the depositional system, from siliciclastic shelf sedimentation, to a 'restricted' basinal environment; conditions that were to persist until late Llanvirn times.

Arenig-Llanvirn Shales; Following the accumulation of the mid-Arenig arenaceous successions, marine mudstone deposition dominated the latter part of the Whitlandian through to the upper parts of the Llanvirn. The entire shale sequence is collectively referred to as the Arenig-Llanvirn Shales. Faunal evidence and lithostratigraphic continuity, locally allows broad divisions to be made, i.e. Penmaen Dewi Formation (Arenig), *D. artus* Shales (Lower Llanvirn), and *D. murchisoni* Shales (Upper Llanvirn). Deposition during this prolonged period of argillic sedimentation was predominantly by suspension and fine grained muddy turbidites, however, depths are not likely to have been much greater than 500m.

Fishguard Bay Group: The Fishguard Bay Group represents the development of late Upper Llanvirn, Llandeilo, and early Caradoc (Constonian Stage) sediments in the Fishguard – Dinas Head region. The Group is divided into three principle stratigraphic units, the Lower Town Formation, Dyffryn Formation and *N. gracilis* Shales (Fig. 4.8.). The Lower Town Formation represents a 'deep' marine clastic sequence, the detrital mineralogy of which suggests that shallow marine conditions were prevailing in the immediate vicinity during late Llanvirn times. The Dyffryn Formation is mudstone dominated and of Lower Llandeilo age. The *N. gracilis* Shales crop out to the east of this area. It is likely that they are the along stike continuation of the anomolously thick Fishguard Bay Group (≈2km).

In conclusion, the unravelling of the sedimentary stratigraphy (Maps 1-4) into major stratigraphic units may go some way to aiding further work on the sedimentary sequence developed within the Fishguard – Porth-gain district. It is apparent from the interpretations presented above, that the nature of sedimentation varied considerably. This is likely to relate to the fact that the sequence records deposition from the Middle Cambrian to the Upper Llandeilo (*c.* 100 m.y.). During which time changes in the style of sedimentation reflect variation within the local depositional system, basin-wide events, and global eustatics.

However, many questions remain largely unanswered. For example, can the sedimentary history recorded here (with no reason to suggest that such complexity it is not applicable elsewhere in SW Dyfed), be accommodated in the present 'classic' model of the Welsh Basin ?, or even within a series of smaller interrelated basins ? Was SW Dyfed consistently situated along the southern margin of a basin situated to the North ? (i.e. what was to the south of Dyfed). This study area is rather 'isolated' in terms of answering such questions, it requiring the sequences both north and south to be investigated, before a fuller understanding of the sedimentary history of SW Dyfed can be achieved and some of the above questions answered.

CHAPTER FIVE
IGNEOUS STRATIGRAPHY

5.1. INTRODUCTION

The Fishguard Volcanic Complex represents the thickest developments of Caledonian volcanics in South Wales, and has proved a target for copious research {1.3}. In the light of recent research (Rowbotham & Bevins (1978), Bevins (1978, 1979, 1982), Bevins & Roach (1979a,b;1982), Bevins *et al.* (1984), Kokelaar *et al.* (1984a,b), Lowman (1977), Lowman & Bloxam (1981)), a detailed study has not been undertaken. The spatial distribution of the principle igneous lithologies west of Fishguard, have been delimited, and much of this information is depicted on Maps (1 to 6). The discontinuous exposure, complex tectonism, and rapid lateral and vertical facies change, only allows the differentiation of the volcano-stratigraphy however, at coastal sections (Maps 5 & 6; see also Kokelaar *et al.* 1984b, Figs. 7,8,9).

In briefly considering the igneous stratigraphy of the area, it is appropriate to discuss the volumetrically minor, although important, range of igneous rock types which are not associated with the Fishguard Volcanic Group. For this reason the following chapter addresses igneous geology under the broad lithologic and stratigraphic headings of: 'Isolated Tuffs', Arenig igneous activity (Pwll Deri Tuffs & Yns Castell Ash), and Llanvirn igneous activity (Fishguard Volcanic Complex).

5.2. 'ISOLATED' TUFFS

Thin (<20cm) tuffs which occur as 'isolated' beds with no apparent source, crop out sporadically within the marine mudstones of the Arenig-Llanvirn Shales of the Fishguard – Porth-gain district. Similar deposits are documented at all biostratigraphic levels from the Arenig to the Llandeilo throughout much of SW Dyfed; variously termed in early publications as *'china-stone'* and *'halliflinta'* (e.g. Part 1922, Williams 1934) and more recently as *'distal tuffs'* (Kokelaar *et al.* 1985). The sources of many such deposits remain, for the most part uncertain, and whilst subaqueously emplaced, there is often little evidence to indicate whether or not they reflect deposits from subaerial or submarine eruptions. This is further complicated, as to whether or not, they represent true volcanogenic deposits sourced from active volcanism (i.e. igni-turbidites, air-fall units, subaqueous ash-falls) or are the product of essentially 'epiclastic' processes and remobilisation of poorly consolidated ash; the distinction between such deposits is often not possible.

Fig 5.1. (A). Principle volcanostratigraphic units and surrounding strata developed within the Fishguard – Porth-gain district. **(B).** Historical division of the Fishguard Volcanic Group.

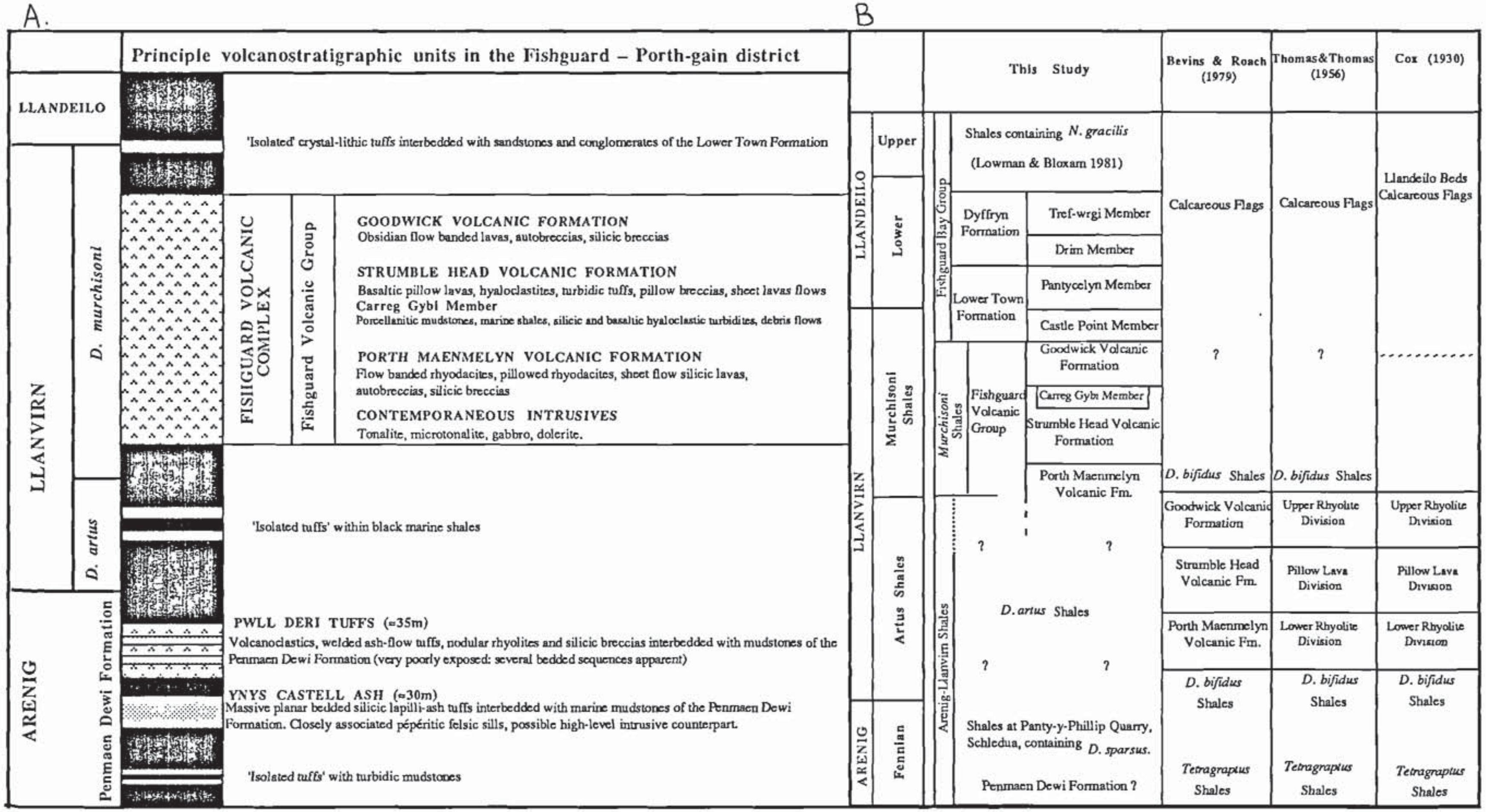
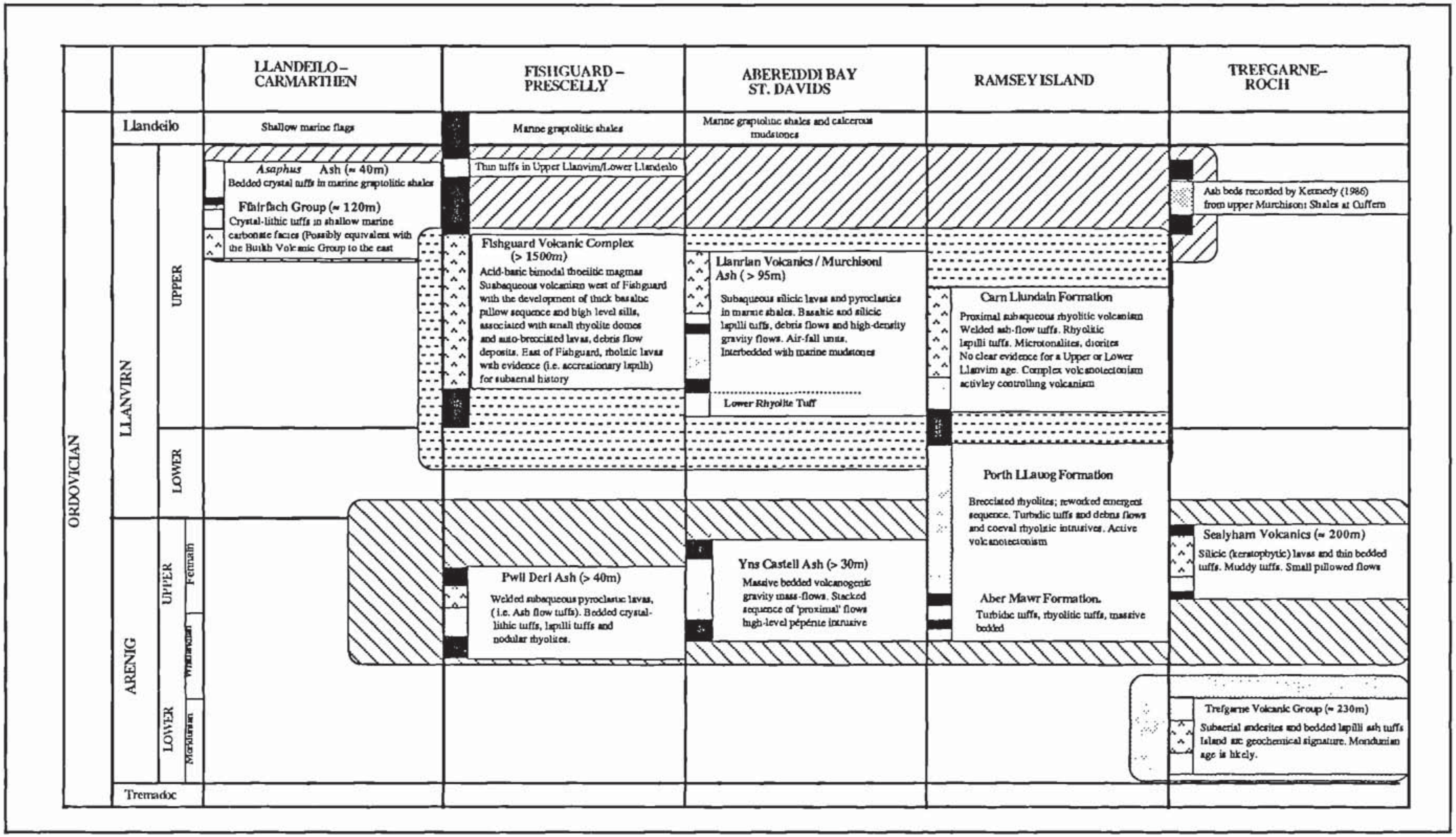

A.
Principle volcanostratigraphic units in the Fishguard – Porth-gain district
LLANDEILO
LLANVIRN
ARENIG
D. murchisoni
D. artus
Penmaen Dewi Formation
'Isolated' crystal-lithic tuffs interbedded with sandstones and conglomerates of the Lower Town Formation
FISHGUARD VOLCANIC COMPLEX
Fishguard Volcanic Group
GOODWICK VOLCANIC FORMATION
Obsidian flow banded lavas, autobreccias, silicic breccias
STRUMBLE HEAD VOLCANIC FORMATION
Basaltic pillow lavas, hyaloclastites, turbidic tuffs, pillow breccias, sheet lavas flows
Carreg Gybl Member
Porcellanitic mudstones, marine shales, silicic and basaltic hyaloclastic turbidites, debris flows
PORTH MAENMELYN VOLCANIC FORMATION
Flow banded rhyodacites, pillowed rhyodacites, sheet flow silicic lavas, autobreccias, silicic breccias
CONTEMPORANEOUS INTRUSIVES
Tonalite, microtonalite, gabbro, dolerite.
'Isolated tuffs' within black marine shales
PWLL DERI TUFFS (≈35m)
Volcanoclastics, welded ash-flow tuffs, nodular rhyolites and silicic breccias interbedded with mudstones of the Penmaen Dewi Formation (very poorly exposed: several bedded sequences apparent)
YNYS CASTELL ASH (≈30m)
Massive planar bedded silicic lapilli-ash tuffs interbedded with marine mudstones of the Penmaen Dewi Formation. Closely associated pépéritic felsic sills, possible high-level intrusive counterpart.
'Isolated tuffs' with turbidic mudstones
B
This Study
Bevins & Roach (1979)
Thomas&Thomas (1956)
Cox (1930)
LLANDEILO
Upper
Lower
LLANVIRN
Murchisoni Shales
Artus Shales
ARENIG
Fennian
Fishguard Bay Group
Shales containing N. gracilis
(Lowman & Bloxam 1981)
Dyffryn Formation
Tref-wrgi Member
Drim Member
Lower Town Formation
Pantycelyn Member
Castle Point Member
Murchisoni Shales
Fishguard Volcanic Group
Goodwick Volcanic Formation
Carreg Gybi Member
Strumble Head Volcanic Formation
Porth Maenmelyn Volcanic Fm.
Arenig-Llanvirn Shales
?
D. artus Shales
Shales at Panty-y-Phillip Quarry, Schleduа, containing D. sparsus.
Penmaen Dewi Formation ?
Calcareous Flags
Llandeilo Beds Calcareous Flags
D. bifidus Shales
Goodwick Volcanic Formation
Upper Rhyolite Division
Strumble Head Volcanic Fm.
Pillow Lava Division
Porth Maenmelyn Volcanic Fm.
Lower Rhyolite Division
D. bifidus Shales
Tetragraptus Shales

Figure 5.2. Regional correlation and principle lithologic features of the main Lower Ordovician volcanic centres throughout SW Dyfed. A speculative corelation into 4 stratigraphic events has been undertaken based on the numerous documented sources the (see Bevins & Roach 1979a, Allen 1982).


ORDOVICIAN
ARENIG
LLANVIRN
Llandeilo
Tremadoc
LOWER
UPPER
Moridunian
Whitlandian
Fennian
LLANDEILO–CARMARTHEN
FISHGUARD–PRESCELLY
ABEREIDDI BAY ST. DAVIDS
RAMSEY ISLAND
TREFGARNE–ROCH
Shallow marine flags
Marine graptolitic shales
Marine graptolitic shales and calcerous mudstones
Asaphus Ash (≈ 40m)
Bedded crystal tuffs in marine graptolitic shales
Ffairfach Group (≈ 120m)
Crystal-lithic tuffs in shallow marine carbonate facies (Possibly equivalent with the Builth Volcanic Group to the east
Thin tuffs in Upper Llanvirn/Lower Llandeilo
Fishguard Volcanic Complex (> 1500m)
Acid-basic bimodal tholeiitic magmas Subaqueous volcanism west of Fishguard with the development of thick basaltic pillow sequence and high level sills, associated with small rhyolite domes and auto-brecciated lavas, debris flow deposits. East of Fishguard, rhyolitic lavas with evidence (i.e. accretionary lapilli) for subaerial history
Llanrian Volcanics / Murchisoni Ash (> 95m)
Subaqueous silicic lavas and pyroclastics in marine shales. Basaltic and silicic lapilli tuffs, debris flows and high-density gravity flows. Air-fall units. Interbedded with marine mudstones
Lower Rhyolite Tuff
Carn Llundain Formation
Proximal subaqueous rhyolitic volcanism Welded ash-flow tuffs. Rhyolitic lapilli tuffs. Microtonalites, diorites No clear evidence for a Upper or Lower Llanvirn age. Complex volcanotectonism activley controlling volcanism
Ash beds recorded by Kennedy (1986) from upper Murchisoni Shales at Cuffern
Porth LLauog Formation
Brecciated rhyolites; reworked emergent sequence. Turbidic tuffs and debris flows and coeval rhyolitic intrusives. Active volcanotectonism
Aber Mawr Formation.
Turbidic tuffs, rhyolitic tuffs, massive bedded
Pwll Deri Ash (> 40m)
Welded subaqueous pyroclastic lavas, (i.e. Ash flow tuffs). Bedded crystal-lithic tuffs, lapilli tuffs and nodular rhyolites.
Yns Castell Ash (> 30m)
Massive bedded volcanogenic gravity mass-flows. Stacked sequence of 'proximal' flows high-level pépérite intrusive
Sealyham Volcanics (≈ 200m)
Silicic (keratophytic) lavas and thin bedded tuffs. Muddy tuffs. Small pillowed flows
Trefgarne Volcanic Group (≈ 230m)
Subaerial andesites and bedded lapilli ash tuffs Island arc geochemical signature. Moridunian age is likely.

Isolated tuffs from the Fishguard – Porth-gain district.

'Isolated-tuffs' occur at numerous localities throughout the Arenig-Llanvirn Shales. Examples of such deposits are well preserved at Aberfelin [S.M. 8320 3240] and Traeth Llfyn [S.M. 8005 3180]. The latter crop out immediately to the south of this study area, north of Abereiddi Bay, underlying the Llanrian Volcanic Group (see Hughes *et al.* 1982, their locality 2D Figure 4.), they represent good examples of this type of deposit and therefore are briefly discussed. The majority of the tuffs occur as individual beds isolated in marine shales, rarely, stacked sequences are developed. The thin bedded nature and fine grain size (mud - silt - fine-sand) suggests that they are essentially 'distal' (*sensus* Kokelaar *et al.* 1985) from source. The deposits show no evidence to suggest an 'air-fall' (or subaqueous ash-fall) origin and are interpreted as density-flows. This is substantiated by their erosive properties and internal sedimentary structures which suggest traction and suspension deposition from fine grained density currents (Plate 5.1 & 5.2.). The origin and phase transformations within high concentration turbidity currents and debris flows (with particular emphasis and application to volcanogenic deposits), has received considerable attention (see Fisher 1983, 1984, and references therein). However, tectonic modification of sedimentary structures within the tuffs thr ughout the Fishguard – Porth-gain district often makes interpretation of the dominant grain s pport mechanism (i.e. turbidity currents (turbulence), grain flow (dispersive pressure or grain to grain interaction), fluidised flow (upward flow of escaping pore fluid), and debris flow (cohesive mud m trix strength) *c.f.* Middleton & Hampton 1976) difficult .

Isol ted tuffs at Aberfelin: The shales which occupy the coastal districts of Aberfelin are be eved to be of Arenig age (i.e. Penmaen Dewi Formation). The entire coastal sequence co ains black, pyritiferous mudstones which may reflect suspension deposition (i.e. 'pel gic-rain'), and silt-laminated mud turbidites {4.6.}. Interbedded with the turbiditic mudstones are several thin bedded tuffs, which are readily identifiable due to their vivid bleached colouration (Plate 5.1.). They are comprised of unresolved fine-sand to silt-grade detritus and angular to subangular feldspar with minor quartz.

Flow description

Individual tuffs (here after referred to as flows) vary in thickness from 5-15cm. Where the effects of tectonism are minimal, flow units show lateral continuity through outcrop. Basal contacts vary from deeply scoured to planar, whilst upper surfaces are sharp to weak undulatory. Internally, individual flows show a variety of structures. However, layering is common to many flow units and suggests therefore, that they were deposited in a similar manner, and by a current; the mechanics of which were probably communal. Whilst all flows show salient similarities, two distinct flow types have been observed: *scoured flows* (Flow A, Plate 5.1.), and *planar bedded flows* (Flow B, Plate 5.1.).

Scoured flows are recognised by their deep eroded base and planar top (Plate 5.1.). The basal contact varies from undulatory, to what can be best termed a 'channelled' surface, the depth of which may locally approach half of the flow thickness. Layering within scoured flows, whilst diffuse, can be discriminated in terms of two layers: a lower 'debris-layer' which is comprised of mudstone intraclasts suspended in silt-grade detritus, which passes upwards into an upper layer of weakly structured silt to fine-sand grade tuff and shows little evidence for obvious size grading (Plate 5.1.). The transition between the two layers is gradational. Fine planar to indistinct laminae may define a diffuse bedding toward the top of the flow. The debris layer may approximate more than 50% of the flow unit, of which, more than 70% by volume may be comprised of mudstone 'rip-ups'. The intraclasts are sub-rounded to angular, and vary in size from small flakes to boulders. All intraclasts show evidence for a stretching fabric and bedding planar imbrication (Plate 5.1.). The intraclasts may show a crude reverse grading.

Planar bedded flows are similar in general character to the scoured flow units, although they lack extensive erosional features. In this respect, it is thought that planar bedded flows may reflect lower velocity, although not necessarily more 'dilute' currents. They may show the presence of a thin structureless siltstone above a weak undulatory to planar base. This passes upwards into an intraclastic mudstone 'debris' layer, similar in character to the scoured flow units, although, intraclasts show evidence for extensive flattening (non-tectonic), exhibited by elongation of intraclasts in the horizontal plane of the flow. This flattening fabric may locally approach the point where intraclasts are 'welded' to form a near continuous mudstone band, ind cating the 'rip-ups' were poorly lithified, and highly susceptible to internal shear during flow deposition. The debris-layer passes abruptly into weak graded tuff which may exhibit pronounced parallel lamination (Plate 5.1.).

Interpretation of scoured flow units

Sc ured flows remain largely enigmatic, in so much as, the incorporation of large mudstone intraclasts from the sea-floor on the scale observed, is not a feature generally associated with thin bedded fine grained volcanogenic (or other) density-flows. The strongly erosive properties would tend to suggest that the original flow was of high-velocity and high-density. The origin for the two-fold layering (debris-layer and structureless silt) remains largely uncertain. However, a possible scenario is that, the scoured flows reflect two phases of deposition: firstly from the head of the density-current and then by the overriding 'tail/body'. A concept, which has similarities to the 'flow-head' deposits which may precede the T_a division of ash-flow turbidites (Sparks & Wilson 1983). Based on this consideration, a possiple explanation may be as follows:

The debris-layer is a single depositional unit. The layer is likely to have formed as the direct result of erosion by the snout (or flow head) of the original density-current, the snout being commonly regarded as the main site within a turbidity or density current where erosion can take place (Allen 1984). It is thought that the flow, whilst assimilating intraclasts into the flow body was modifying itself, in so much as, it would have changed from an original silt grade density-flow (as observed from the matrix of the debris-layer) into one in which a large grain size distribution would be apparent. In short, the flow, by virtue its own erosive properties, became a coarse grained gravity flow (i.e. coarse-grained turbidite, or debrite). The shale blocks may have increased the dispersive properties sufficiently so as to behave in a manner similar to a rigid plug, at which point, clasts were supported by the cohesiveness of the sediment-water mixture (i.e. debris flow) with deposition of the debris-layer occurring by cohesive freezing (Lowe 1982, Vesser 1983). The 'body/tail' of the density-current would override the rigid plug, (and may possibly rework its upper part) with subsequent deposition by normal current processes (i.e. traction and suspension), thus forming the observed two fold la ring. If correct, the interpretation of a flow modifying itself by its own erosive properties wi h entrainment of lithic blocks creating a 'self-induced' rigid plug, is a mechanism of flow-modification and transformation, not widely discussed in the literature (see Naylor 1980). However, why the flows w ere so erosive in the first place is uncertain. It may possibly reflect bo y transformation (*c.f.* Fisher 1983) within the original density current, related to transportation down a depositional slope with a change from laminar to turbulent flow. In this respect it is of interest to note that in the vicinity of Aberfelin, 'soft-sediment' folds and 'bal ed-up' sandstones are present within the marine shales, which may speculatively suggest th the environment was one of a slope facies.

Isolated tuffs at Traeth Llfyn (D. artus Shales): Within the *D. artus* Shales (Aber Mawr Shales of Hughes *et al.* 1982), underlying the Llanrian Volcanic Group at the south of Traeth Llfyn [S.M. 8012 3180], thin tuffs crop out on the small wave polished platform. The sequence at this locality is inverted and the observed structures and depositional sequence are the wrong way up when viewed directly at outcrop.

It is thought that the tuffs at this locality are best interpreted in terms of deposition from turbidity currents. Individual flows range from 5-12cm thick and are laterally continuous through outcrop. They are readily recognised within the marine shales due to their bleached colouration (Plate 5.2.). The flows are composed of fine silt to fine-sand grade tuffaceous material, rare subangular feldspars are seen in thin section (amorphous prehnite is also present). The lower contact, are invariably undulatory, commonly possessing evidence for load and injection-like structures, a feature typical of fine grained turbidites (Stow & Piper 1984). The flame and injection structures may relate to subsequent overburden pressure on the poorly lithified bed beneath the flow unit, although, deformation may also have arisen by shearing on the substratum by the overriding flow. Layering and structures within individual flows are diffuse, although, a common lower division of convoluted fine, silt-grade ripples indicates periods of traction deposition. Small (2mm to 3cm) bladed and flattened mudstone 'rip-up' clasts are readily identified. Intraclast rounding coupled with their orientation – imbrication parallel to the planar bedding (Plate 5.2) may support the suggesting of traction current deposition for the lowermost parts of the flows (Vesser 1983). Weak grading and lamination present in the upper parts are thought to reflect deposition as a direct result of suspension fall-out. In short the tuffs, represent deposition from high-density silt-grade turbidity currents.

Hughes *et al.* (1982) document tuffs from the sequence at Traeth Llfyn which they suggest are both density flows and 'air-fall' units. The present author believes that whilst density flows can be readily distinguished (Plate 5.2), the discrimination of 'air-fall units', as suggested by Hughes *et al.* (1982) on the basis of sharp non-erosive lower contacts, lack of mudstone intraclasts and gradational upper surfaces, fails to be conclusive, in so much as, such features may also be characteristic of turbidity current deposition. However, if an air-fall (or subaqueous ash-fall) origin is appropriate, then the sequence at Traeth Llfyn exhibits the interaction of 'distal' volcanogenic processes that were operative in SW Dyfed throughout much of the late Arenig and Llanvirn. The thin density-flows may reflect remobilisation of unconsolidated ash from a site of active volcanism, or the 'distal-parts' of proximal debris flows or pyroclastic flows which developed around the sites of active extrusion. The 'fall' units would reflect the direct influence of volcanism, although the nature of such volcanism and position of the centre is not known. Also, in the case of the gravity flows, a source may have been 10's of km from their present position (i.e. Yamada 1973, 1984, Niem 1977). In this respect, the tuffs have little regional stratigraphic significance, although, their apparent presence at all biostratigraphic levels suggests that distal volcanism was an important source of detritus into the sedimentary environment during late Arenig-Llanvirn times.

5.3. ARENIG IGNEOUS ACTIVITY

The earliest recorded volcanism in SW Dyfed is provided by the Trefgarne Volcanic Group (Fig 5.2.). The Group is isolated, in so much as, it has its own geochemical signature (island-arc volcanism, Bevins *et al.* 1984), lithological characteristics (andesitic lavas and tuffs), and stratigraphic position within the lowermost Arenig (Moridunian Stage, Fortey & Owens 1987). However, there appears to be sufficient documented evidence, to suggest, that a

Plate 5.1. 'Isolated' tuffs from the Penmaen Dewi Formation (Upper Arenig) at Aberfelin [S.M. 8320 3240]. Note deeply scoured (Channelled) lower surface to 'Flow A' , obscured in part due to the volumetric abundance of mudstone 'rip-ups'. The flow is thought to reflect a high-density/high-velocity gravity flow current, which may have modified its flow characteristics by the assimilation of lithic's into the flow body. Note stretching fabric to intraclasts and a weak reverse grading in intraclast size. Flow B is a planar bedded flow unit which shows similarities to the scoured flow, in so much as it contains a 'debris-layer', although here mudstone intraclasts are almost 'welded' together to form a near continuous mudstone band. Note coarse parallel lamination at the top of the flow.

Plate 5.2. Isolated volcanogenic tuffs from the *D. artus* Shales (Lower Llanvirn) at Traeth Llfyn [S.M. 8005 3180 : south of this study area]. Note flame structures along the base of the flow and the fine mudstone clasts within the flow body (*n.b.* the shale sequence/flow units are inverted at this locality and depositional features are the wrong way-up, as viewed directly at outcrop)

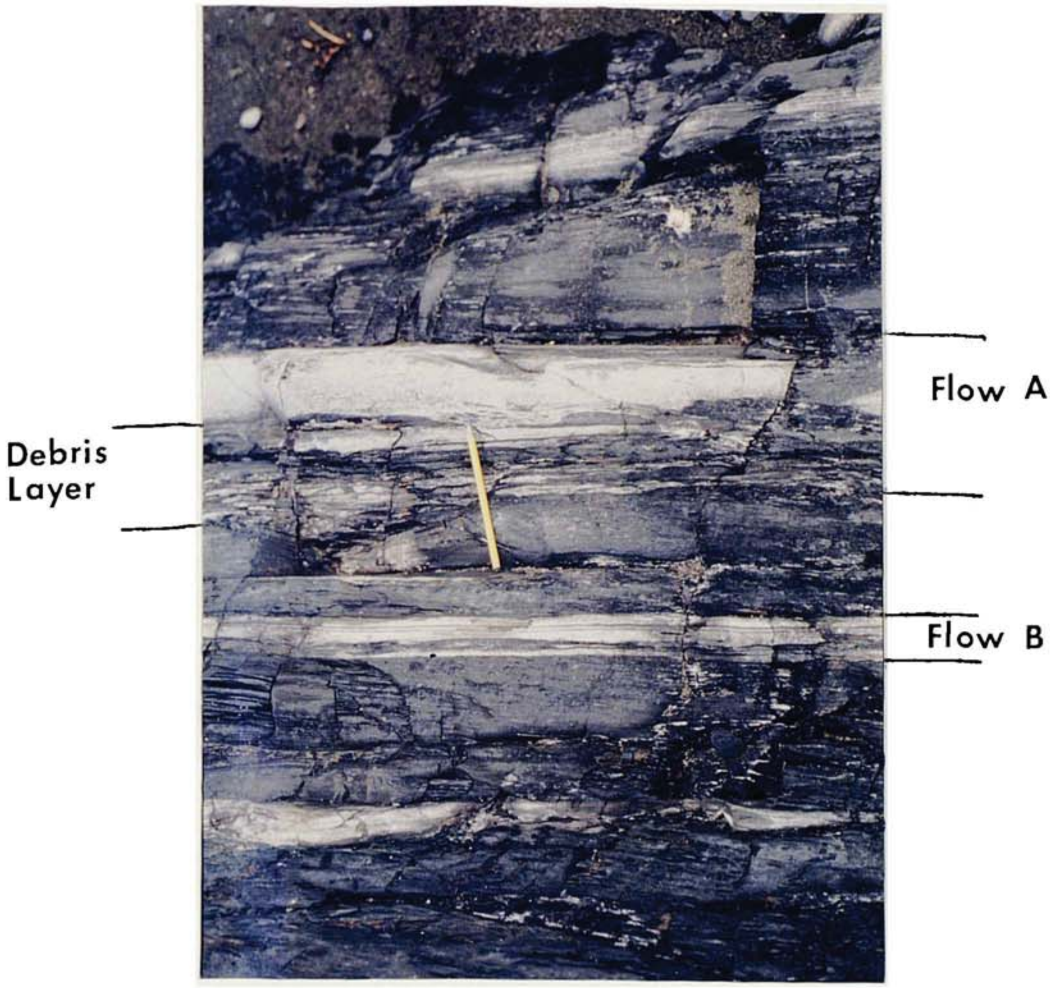

Flow A
Debris
Layer
Flow B

Plate 5.3. Welded ash–flow tuff from the Pwll Deri Tuff. This field photograph is taken of a block which has fallen from its *in situ* position from the cliff section above. The wave polished surface enhances structures which are diffuse when seen at outcrop. The streaked nature of the lithology results from the preferential weathering of chloritised collapsed fiammé. Similar lithologies are seen *in situ*. In thin section textures are invariably obscured by the low-grade alteration.

Plate 5.4. Pépérite breccia from the wave-cut platform at Pwllstrodur [S.M. 8660 3380]. The chaotic pépérite forms part of a continuous *in situ* autobrecciated sill which can be intermittently traced to Pwll Llong where the sill is emplaced within Arenig shales of the Penmaen Dewi Formation. Note rounded terminations which may reflect original outer lobate protrusion surfaces. Many angular fragments can be fitted back together. Several semi-circular enclaves are seen which may represent autobrecciated terminations of small

further, and possibly more widespread episode of Arenig volcanism, affected much of SW Dyfed during late Whitlandian-Fennian times, and possibly spanning the Arenig/Llanvirn boundary (Fig. 5.2.).

In the Fishguard – Porth-gain district, whilst evidence for a volcanic source is lacking, this episode (or episodes) is represented by the Yns-Castell Ash and Pwll Deri Tuff (Fig. 5.1.). The exact biostratigraphic range of both successions is uncertain, however, they occupy a similar stratigraphic position within shales of the Penmaen Dewi Formation (i.e. Arenig mudstones) and are realistically assigned a late Whitlandian/Fennian age. Further evidence for 'late Arenig' igneous activity within the area, comes in the form of thick autobrecciated sills. These were emplaced into mudstones of the Penmaen Dewi Formation whilst they were semi-lithified. It is thought that the sills may represent the remnant hypabyssal counterparts of the Arenig extrusives.

(a). Yns-Castell Ash (YCA); *Type locality [S.M. 8510 3385]; Thickness ≈ 30m.*

The Yns-Castell Ash (*c.f.* Cox 1930) represent a thick (≈ 30m) bedded sequence of silicic lapilli-ash tuffs. The tuffs are best observed on the headland of Yns y Castell at the entrance of Abercastle Harbour [S.M. 8510 3385] where they can be seen to lie with concordance on shales of the Penmaen Dewi Formation (i.e. Arenig). The suggestion that the shales are of Arenig age (and therefore, that the tuffs are also of Arenig age) is based on the recognition that the laterally e uivalent shales associated with the Yns Castell Ash at Pwll Whitting (Map 4.), have yielded xtensiform graptolites (Cox 1930). The age of the overlying shales is uncertain, although they are also realistically assumed to be of Arenig age also. At Yns y Castell the tuffs can be seen to be flat lying. Broad undulatory folding of the sequence towards Ynys Deullyn, reflects the r lative competence of the unit within the host shales. The tuffs can be traced along the coastline westwards to Pwll Whitting, although here, the succession is thrust and block-faulted (Map 4.).

Description

The tuffs form stacked planar bedded units, varying in thickness from less than 1m to ≈ 5m. Individual beds show a tendency to become progressively thinner towards the top of the succession. Each bed is readily defined by a sharp to weak undulatory base. Twenty-eight individual beds can be recognised at Yns y Castell, with little or no evidence for intervening mudstone (i.e. back-ground sedimentation) between each bedded unit. Shale 'rip-ups' from the underlying Penmaen Dewi Formation can be seen within the lowermost bed, suggesting that the tuffs represent the product of mass-flow deposition. Internally, individual beds show limited evidence for sorting and are generally massive in appearance. A diffuse normal grading is locally apparent, although no discrete coarse tail-grading has been observed, a feature common to similar deposits. Angular to sub-rounded, weakly porphyritic and non-vesicular and scoriaceous lapilli to of block size clasts are a common feature. However, the primary matrix mineralogy is altered by the effects of low-grade metamorphism with the original glassy material now converted to chlorite. The top of each bed is defined by a thin weakly laminated fine to coarse grade tuff, which sharply defines the metre-scale bedding.

Interpretation

The 'Yns Castell Ashes' are interpreted as a series of subaqueously deposited volcanoclastic density-flows. It is not certain whether they represent deposition as a direct result of volcanism (i.e. pyroclastic flows, *sensus* Fisher & Schmincke 1984) or the products of remobilisation of poorly consolidated ash. A direct pyroclastic origin is favoured here. Evidence to support a direct pyroclastic origin is suggested by the lack of intervening marine sediments, probably reflecting relatively rapid and near continuous deposition related to successive, although waning eruptions (i.e. beds become thinner towards the top). The absence of current or wave reworking indicates that the beds where deposited significantly below effective storm wave base. Poor sorting and weak undulatory and erosive lower surfaces suggest flows characteristics similar to high-concentration coarse grained turbidity currents. Poor sorting and lack of grading is character with 'proximal' pyroclastic flows, in so much as prolonged transportation will lead to flows becoming density and size graded, vertically as well as laterally (Yamada 1984). The thin laminated coarse to fine ash layer which defines individual flow tops is thought to represent fine fall-deposits reflecting the settling out of fine material that was placed in suspension as a turbulent cloud during deposition. Such units may equally represent a combination of, suspension fall-out, and less dense turbidity flows which separated from and lagged behind the denser current (Yamada 1984) The poor sorting and occurrence of both non-vesicular and scoriaceous lapilli and blocks size clasts may suggest that the tuffs originated from hydroclastic eruptions (Fisher & Schmincke 1984) above the Pressure Compensation Level (PCL Fisher 1984) for silicic volcanism; the depth (pressure) of which is is dependant on many variables, although may be in the order of ?500-1000m (see Fisher 1984, Kokelaar *et al.* 1985). A source area is lacking for the deposits, although their proximal characteristics, and the presence of true subaqueous ash-flow tuffs (welded) at a comparable stratigraphic level within the Pwll Deri Tuff to the east (see below), indictes that they may represent proximal pyroclastic deposits from a volcanic source in the immediate vicinity of the Fishguard region. Equally, a small volcanic centre may lie hidden within the inland sequence.

(b). Pwll Deri Tuff (PDA); *Type Locality [S.M. 8915 3825]; Thickness (> 35m).*

The Pwll Deri Tuff is a newly recognised sequence of welded ash-flow tuffs, pyroclastics, and volcanoclastics. The *Tuff* crops out discontinuously in the vicinity of Pwll Deri [S.M. 8915 3825] and is very poorly exposed. At Pwll Deri, the presence of a silicic volcanic sequnce can only be implied from loose boulders at the base of the cliff on the south side of the bay. Accessible exposure to the *Tuff*, is restricted to the steep cliffs at the south of Pwll Deri, several small exposures along the coastal path [S.M. 8925 3830] and two small quarries within the cliff section which have be excavated specifically for the volcanics. The *Tuff* sequence is interbedded with shales from which Cox (1930) records the presence of extensiform graptolites, indicating that an Arenig age is appropriate. Such an age is supported, by their spatial attitude at outcrop, above the mid-Arenig shallow marine sediments of the Trwyn Llwyd Formation (Maps 2 & 3).

At outcrop, four individual volcanogenic horizons (≈ 5-10m thick) (each of which separate by shales), can be traced sporadically, (Map 3.). In this respect the Tuff differs from the Yns-Castell Ash to the east, in so much as, periods of marine sedimentation punctuate discrete events. However, both sequences occupy a comparable stratigraphic level, and it is likely that they represent part of the same event, or episode of events. The lithological nature of the *Tuff* can be observed directly at outcrop, although a far greater (and less hazardous) appreciation of the various rock types can be gained from large blocks at the foot of the cliffs south of Pwll Deri. Here, textures and structures are enhanced by wave polishing (Plate 5.3.). The variety of lithologies observed, include; welded and non-welded pyroclastic ash-flow tuff, crystal-lithic tuffs, nodular rhyolites and various silicic breccias. The presence of welded ash-flow tuffs, indicate that true pyroclastic lavas constitute a significant part of the sequence. Whether or not they represent the product of subaqueous or subaerial eruption remains uncertain. Welded pyroclastic flow, have proved to be controversial when interpreted to be subaqueously erupted, or emplaced subaqueously from subaerial vents on stratigraphic grounds (Fisher & Schmincke 1984, Cas & Wright 1986). Whilst their is no evidence for the source of the deposits, it is contended here, that the welded ash-flow tuffs of the Pwll Deri Tuff represent at least emplacement, if not eruption (see Kokelaar *et al.* 1985), within a marine environment below effective storm wave base.

The presence of true pyroclastic lavas may suggest that the Fishguard region was one of possibly several active volcanic centers during the upper parts of the Arenig. Such a volcanic centre may have shed material into the surrounding areas resulting in the common occurrence of 'distal turbidic tuffs' as seen within many of the Arenig shale sequences throughout SW Dyfed. More proximal deposits being represented by sequences such as the Yns Castell Ash. In summary, the Pwll Deri Tuff can be best described as an interbedded sequence of subaqueous crystal-lithic tuffs, welded and non-welded ash flow tuffs, and nodular and brecciated rhyolites.

c). Arenig intrusive activity [BF on Maps 3&4, Sill complex between Pwllstrodur and Pwll Llong]

The timing of intrusive activity throughout much of SW Dyfed is rightly assumed to be of Llanvirn age, coeval with the major phases of volcanism. However, in the light of Arenig extrusive activity above, and evidence documented elsewhere in SW Dyfed (Kokelaar *et al.* 1985), the possiblity arises that all the intrusives may not be confined to the Llanvirn. Using field relationships and the well documented phenomenon of magma – wet-sediment interaction (see Kokelaar 1982, Hanson & Schweickert 1982, Lorenz 1984), with its inference of penecontemporanueous emplacement with sedimentation, it is possible to suggest that at least some of the intrusives in the Fishguard – Porth-gain district are of Arenig age.

Based on the phenomenon of wet sediment/magma interactions, an example of an Arenig intrusive, crops outs intermittently between the headland north of Pwll Llong [S.M. 8405 3340] and Pwllstrodur (Maps 3 & 4, denoted as BF). The Arenig age of the host shale is proven by the recorded occurrence of the extensiform graptolites *D. nitidus* Hall and *D.* cf. *extensus*, *D.* cf. *sparsus* (Cox 1916). The penecontemporanueous emplacement of magma is indicted by the extensive development of pépérite in off-shoots and adjacent to sill surfaces, and magma-sediment relationships (i.e. *in situ* auto-brecciation and hyaloclastisation). As outlined by Hanson & Schweickert (1982), pépérites/pépérite breccia, result from the interaction of magma or lava, chilling and undergoing fragmentation upon injection or emplacement into poorly lithified or unconsolidated sediment. The resulting deposit is one of chaotic intermixed igneous and sedimentary material.

Description of Pépéritic intrusive from Pwllstrodur

The sill in question (BF on Map 3&4) can be traced from Pwll Llong to Abercastle, to the small wave-cut platform at the entrance to Pwllstrodur [S.M. 8660 3380]. Evidence for *in situ* auto brecciation is evident at all these localities. However, the pépérite exposed at Pwllstrodur is the most informative, and sediment/magma relationships are highlighted by wave polishing. The lithology has been previously documented by Cox (1930, p.331), as the *'Pwll-Strodyr Fault-Intrusion (felsite)'*, its origin he interpreted as felsite emplacement during periods of active faulting.

The pépérite unit forms part of a highly altered sill of seemingly intermediate composition, which is intruded into weakly laminated black shales. The intermediate character of the sill, is suggested by the presence (in thin section) of abundant tracyhtiod feldspar phenocrysts, rare psuedomorphed mafic phenocryst (? hornblende phenocrysts), set within a recrystallised fine grained quartz-feldspathic matrix. The base of the sill is readily discriminated in the northerly facing cliffs at the entrance to Pwllstrodur [S.M. 8662 3377]. The top of the sill can be observed further along the cliff-section [S.M. 8660 3376], although it is poorly exposed and deeply weathered. Where observed, the basal sill-sediment contact is irregular, with lobate protrusions marking the magma-sediment interface, accompanied by localised flaming of sediment into the sill body. A small pépérite lines the lowermost sill surface, which shows clear evidence for *in situ* auto-brecciation and hyaloclastisation. This pépérite is, however, small in comparison to that observed on the wave-cut platform.

As a result of beach gravel cover, the true relationship of the pépérite on the wave-cut platform to the sill in the cliff section is uncertain, although their is no doubt as to them being part of the same intrusive body. It is thought that the deposit may represent an off-shoot from the main sill body. The pépérite unit trends east-west, the overall strike length of which is greater than 25m and locally thickness is in excess of 6m. The pépérite is composed of angular to sub-angular clasts, dispersed in a fine grained matrix of argillic sediment and hyaloclastite. Individual clasts are up to 1.5m in diameter, show abundant evidence for *in situ* auto-brecciation and fragments that can be commonly fitted back together, with mudstone and hyaloclastite debris invading along angular fractured surfaces (Plate 5.4.). Sub-rounded surfaces are evident and thought to represent original outer lobate protrusion surfaces. The presence of rounded and matrix infilled semicircular enclaves are thought to represent intrusive terminations, resulting from digital emplacement in an almost ameboidal-like fashion.

The spatial distribution of clasts is chaotic. Plate 5.4. shows characteristic clast concentration. However, clasts can be observed 'suspended' in shale, up to 3m from the main pépérite unit. This would suggest that the original mudstones behaved as a thixotropic medium, allowing gravitational 'sinking' during and possibly after intrusion. The lack of sedimentary structures within the shales in the immediate vicinity of the pépérite, may indicate total homogenisation, loss of cohesion, and intense 'soft-sediment deformation'. It is also possible, that small scale phreatic explosions and effects such as transplosion as described by Mills (1984) from pillow lavas, created 'seismic shock' aiding liquification of the host sediment and allowing dispersion of clasts.

Summary: In summary, the pépérite breccia accompanying a felsic sill which is intruded into Arenig shales at Pwllstrodur, Abercastle and Pwll Llong suggests emplacement at a time when the sediments were poorly consolidated. It is thought that such intrusives may reflect the hypabyssal counterparts of the Arenig extrusives. The deposits appear typical in many respects to similar pépérites documented in the literature (for details see Kokelaar 1982), being an *in situ* autobrecciated magma which interacted with poorly lithified sediment.

Comparable Late Arenig/Lower Llanvirn igneous activity

The recent detailed studies on Ramsey Island by Kokelaar *et al.* (1984.a., 1985) indicate, that whilst this area was a complex volcanic centre during the Llanvirn, juvenile 'distal turbidic tuffs' (their, Lower Aber Mawr Tuff Member) of Arenig age were deposited in moderately deep water. Elsewhere, whilst uncertainty exists as to the exact biostratigraphic age, several volcanic horizons occur at 'comparable' levels. For example, the Sealyham Volcanic Group is assigned to the upper Arenig - lower Llanvirn (Part 1922, Thomas & Cox 1924, Evans 1945). A further succession of volcanoclastics, termed the Foel Tyrch Beds crop out within the Prescelly Hills (Evans 1944), appearing to occupy a stratigraphic position at either the top of the Arenig or the base of the Llanvirn. Evans (1944) correlates the Foel Tyrch Beds with the Sealyham Volcanic Group. Williams (1934) records a thick ash horizon (*c.* 30m), his 'Upper Llandeloy Ash', within sediments of mid-Arenig age from the Tremaenhir district.

Taking into consideration that the exact biostratigraphic range of succession is uncertain, it seems likely, that the Yns Castell Ash, Pwll Deri Tuff, Sealyham Volcanics and other smaller successions, represent fragmentary evidence for a regional and possibly prolonged late Arenig (early Llanvirn) rhyolitic volcanic episode in this part of SW Dyfed The occurrence within the Fishguard – Porth-gain district of intrusions, the nature of which suggests emplacement in to poorly lithified sediments of Arenig age, supports the contention for late Arenig igneous activity.

5.4. LLANVIRN IGNEOUS ACTIVITY (Fishguard Volcanic Complex)

The Fishguard/Prescelly district records the thickest (>1500m) development of Ordovician igneous rocks in S Wales, referred to collectively as the Fishguard Volcanic Complex. The Complex crops out in a continuous E-W trending belt (>25 km) from Porth Maenmelyn in the west to the Prescelly Hills in the east (Fig. 1.2.). The volcanic successions to the west of Fishguard (this study area), referred to as the Fishguard Volcanic Group (*c.f.* Bevins & Roach 1979a), are dominated by thick basaltic and silicic volcanics of various affinity erupted within a subaqueous environment. The successions to the east of Fishguard are predominantly silicic and whilst subaqueously emplaced, show evidence, such as accretionary lapilli (Lowman & Bloxam 1981) which suggests a partial subaerial history. In this respect it would appear that the Fishguard Volcanic Complex represented both subaerial and subaqueous volcanism. However, the location the position of a subaerial terrain has not been identified, it would have had a limited preservation potential (see Francis 1983) particularly in the light of the erosive nature of the overlying Lower Town Formation {4.6.}, which probably records the unroofing of the Complex within a shallow marine environment.

Age of the Fishguard Volcanic Complex: Based on the early suggestion by Cox (1930) that shales underlying the Porth Maenmelyn Volcanic Formation (Map 5.) were likely to belong to the *D. artus* Shales (= *D. bifidus* Shales), the Fishguard Volcanic Complex has long been regarded as of Lower Llanvirn age {2.4.}. However, no fauna has been collected from Porth Maenmelyn and the age of the shale sequence at this point remains uncertain. To the east, Lowman & Bloxam (1981) identified (from shales within the volcanic successions [S.M. 9635 3770]), a graptolite fauna containing *D. murchisoni, Diplograptus* cf. *coelatus, Glyptograptus shaefei, Glyptograptus* aff. *dentatus*, and *D. artus*, an assemblage indicative of the lower *murchisoni* Biozone. Recently, a similar lower *murchisoni* Biozone, fauna containing *D. murchisoni, D. artus* and *Glyptograptus latus* (Bulman), has been recorded within shales at Tower Hill [S.M. 9615 3700] which immediately underlie the Volcanic Complex (Kennedy 1986). Both assemblages suggest that the age of the Fishguard Volcanic Complex is likely to be Upper Llanvirn (lower part), rather than Lower Llanvirn (. Fig 5.1.).

The Fishguard Volcanic Complex has been correlated with the Llanrian Volcanics at Abereiddi Bay which were thought to be of Lower Llanvirn age (Cox 1916, George 1970, Black *et al.* 1971), although shown more recently to overlie shales which contain a *murchisoni* type fauna (Hughes *et al.* 1982). It seems probable therefore, that the Fishguard Volcanic Complex and Llanrian Volcanics occupy a similar stratigraphic position, significantly above the base of the Llanvirn, and probably restricted to the lower part of the Upper Llanvirn. It is of interest to note that the Builth Volcanic Group to the north-east of Llandeilo appears to be restricted to the Upper Llanvirn (Allen 1982). This suggests that the southern margin of the Welsh Basin was a major centre of volcanic and intrusive activity during the Upper Llanvirn, rather than spread throughout the entire Llanvirn.

5.4.1. FISHGUARD VOLCANIC GROUP

As outlined previously {3.3.}, the Fishguard Volcanic Group spatially dominates the area west of Fishguard (Maps 1&2). Subdivision of the Group as proposed by Bevins & Roach (1979b; i.e. Porth Maenmelyn Volcanic Formation, Strumble Head Volcanic Formation, and Goodwick Harbour Volcanic Formation) whilst not readily distinguished due to tectonism and poor exposure, has been maintained (Fig. 5.1.). One addition to the Strumble Head Volcanic Formation, is the introduction of the term Carreg Gybi Member (the *'Strumble Head Series'* of Thomas & Thomas 1956), to incorporate a thick succession of marine shales, silicic volcaniclastics, and pyroclastics that crop out towards the top of the Volcanic Formation (Map 6). The relationships of the Fishguard Volcanic Group, associated sediments and intrusives, are show on Maps (1-3). Differentiation of the type sections for the Carreg Gybi Member and the Porth Maenmelyn Volcanic Formation is show on Maps (5 & 6). Kokelaar *et al.* (1984b) recently detailed the volcanostratigraphy of the Porth Maenmelyn Volcanic Formation and Goodwick Volcanic Formation at their type localities.

Regional volcanostratigraphy

The regional stratigraphy of the Fishguard Volcanic Group remains poorly appreciated inland, especially, with regard to the local facies variations and the relationships with surrounding strata (Maps 1-3). Field evidence, particularly south of the Pen Caer Peninsular, suggests that the stratigraphy is difficult to interpret in terms of a continuous tripartite succession (Fig. 1.3.). However, allowing for inconsistencies, a tripartite stratigraphy at Formational level is the most effective way to discuss the Volcanic Group with the available outcrop. To fragment the Group into small fault 'isolated' lithostratigraphic units, would at present serve little useful purpose. The repeat successions identified in the volcanostratigraphy at Porth Maenmelyn and Carreg Gybi (Maps 5&6), the lack of evidence for a 'layer-cake' stratigraphy, and vertical and lateral facies change over short distances, indicate that this would also be impractical. This is not to suggest that the sequence is correct and that further modification can not be made, merely it is the simplest and most practical solution with the available data to date.

In the light of the recent research undertaken on the Fishguard Volcanic Group {1.3.}, the following sections discusses the intrusives of the area and the basaltic volcanics of the Strumble Head Volcanic Formation. The author refers you to the recent research (see below) regarding the silicic volcanics (i.e. Porth Maenmelyn and Goodwick Volcanic Formation's) of the Volcanic Group which is the most comprehensive to date.

5.4.2.1. Silicic Volcanism (Goodwick and Porth Maenmelyn Volcanic Formation's)

The subaqueous silicic volcanics of the Goodwick Volcanic Formation and Porth Maenmelyn Volcanic Formation, contain a variety of rock types, and order of lithologies. They exhibit a complex interplay and association of coherent flows, *in situ* auto-breccias, and gravity driven mass-flow deposits, emplaced within a subaqueous environment. There is little evidence to indicate explosive activity. The majority of coherent flows were typically short, thick, flow-banded obsidian lavas characteristic of effusive eruption. An array of breccias and rhyolitic clastic facies rock types developed in response to autoclastic and mass flow mechanisms, rather than large scale phreatomagmatic explosions (Bevins 1982, Kokelaar et al. 1984a,b).

The volcanostratigraphy and geology of both the Porth Maenmelyn and Goodwick Volcanic Formations at their type localities has recently been figured and described by Bevins (1979, 1982) Bevins & Roach (1979a,b, 1982) Kokelaar *et al.* (1984b). Map (5.), extends the Porth Maenmelyn to incorporate the entire sequence between Pwll Deri and Pen Bush, and highlights the repetition in the volcanostratigraphy.

(a). The Goodwick Volcanic Formation (GVF). *Type Locality (S.M. 9450 4050)*
Kokelaar *et al.* (1984b) described and interpreted the Goodwick Volcanic Formation as a sequence of small rhyolite domes, with cores composed of coherent obsidian flows and marginal autobreccias; with gravity flow rhyolite breccias forming in response to instabilities on the slopes.

With reference to the Formation at its type locality as described by Kokelaar *et al.* (1984b), the dolerite which occupies the Pen Anglus headland (Map 2), is thought best interpreted as thrust into complex juxtaposition, rather than a penecontemporanueous slide (Kokelaar *et al.* 1984b, their Figure 8). A further thrust is recognised at the extreme north of the Pen Anglus headland [S.M. 9458 4052]. This brings extrusive basaltic pillow lavas and tube networks onto the dolerite and is not regarded as the pillowed top to the dolerite as suggested by the above workers. As a lithostratigraphic unit, the Formation can be traced inland over a considerable distance towards Carne-Coch [S.M. 9280 3720] although relationships further to the south become obscured by lack of exposure and tectonism. There is little evidence to indicate the continuation of the Formation from Pen Caer to Fishguard.

b) Porth Maenmelyn Volcanic Formation (PMVF). *Type locality [S.M. 8890 3865] .*
The Porth Maenmelyn Volcanic Formation, represents a thick subaqueous accumulation of, bedded silicic tuffs, rhyolitic clastics, coherent flows and intrusives (Maps 1-3 & 5.). The lithotype section on the north side of Porth Maenmelyn, has recently been documented by Bevins & Roach (1979a,b,1982) and Kokelaar *et al.* (1984b). Map (5) details the volcanostratigraphy between Pwll Deri and Pen Bush [S.M. 8805 3965], which incorporates the lithotype section and documents its development to the south of Porth Maenmelyn towards Aber Twn [S.M. 8890 3865]. To the south of the Pen Caer Peninsular, the distribution and relations of the various volcanic successions become obscure due to poor exposure and undoubted structural complexity. Lithologies referred to as the Porth Maenmelyn Volcanic Formation are so named, for their present spatial attitude below basaltic volcanics (i.e. Strumble Head Volcanic Formation). There is little evidence to suggest that the silicic volcanics from the lithotype section, pass conformably from Porth Maenmelyn in the west, to Fishguard in the east.

Type-section (Map 5).
The stratigraphy and lithologies depicted on Map (5) at the type-section are, for the most part consistent with that of previous workers. Apart from the addition of structure and a slightly

more varied range of lithologies, (e.g. tonalites, dolerites, bedded crystal-lithic tuffs), the principle lithologies such as the pillowed rhyodacite lavas (Bevins & Roach 1979b), and obsidian flow banded lava, and rhyolite breccias (Kokelaar *et al.* 1984b) remain largely unchanged. A new rock type is added to the suite of lithologies in the Formation, referred to here as silicic sheet-flow lavas (see below; S on Map 5.; Plate5.5. & 5.6.).

At the type locality, the relationship of the Porth Maenmelyn Volcanic Formation with the Strumble Head Volcanic Formation is regarded as tectonic. Where visible, the contact between both Formations is highlighted by a cataclastite breccia of variable thickness (*c.* 15cm - 1m). The fault has thrust properties, although the displacement and significance of the structure is largely unknown. The cataclastite breccia may only reflect modification of a pre-existing silicic breccia. The contact itself has an undulatory almost stepped profile cutting up section and causing omission with in the upper parts of the volcanostratigraphy. The stratigraphy of the entire coastline is thought to reflect deformation in imbricated fashion, with three repeated units being identified; principally on the basis of a shale-tonalite-dolerite-volcanic association. The entire sequence may be imbricated on basaltic pillow breccias which form the floor of the bay at Yns Y Ddinas [S.M. 8880 3875]. However, the stratigraphy is not 'layer-cake', and facies variations are common over short distances. The complexity observed on the coast, is present to the east, where the repetition of intrusive, sedimentary, and volcanic successions at the coastline is seen in the vicinity of Gawr Fawr [S.M. 8960 3886] and Garn Fechan [S.M. 9000 3890 Map 2.]. Cross-faults, complicate previously thrust units, in part causing the variable outcrop thickness and patterns. Variable thicknesses may also reflect rapid lateral facies change.

'Sheet flow' silicic lavas

Sheet flow silicic lavas are a newly recognised lava type to the Fishguard Volcanic Complex. They lavas occupy the headland of Dinas Mawr [S.M. 8880 3865] which has previously been documented as being comprised of microtonalite sill exhibiting good columnar jointing (Bevins & Roach 1982). Whilst a microtonalite dyke can be seen on the northern most part of the headland (Map 5.), the southern half of the headland is occupied by coherent lavas, the thickness of which coincidentally approximates that of the columnar joints of the microtonalite (Plate. 5.5. .). The lava flows are stacked, and laterally continuous 'sheet-like' flows, which can be readily traced onto the islands of Ynys-y-Ddinas, Ynys Melyn and Carreg Ddu to the west. The lavas at outcrop are variable in thickness from 0.5 – 1.2m, and are characterised by their planar flow habit and their two dimensional aspect ratio which locally approaches 1:750. This would suggest an extremely fluid lava-type. Petrographic evidence supports the contention that these deposits resulted from hot coherent silicic magmas. Feldspar phenocrysts and interlocking microphenocrysts with dendritic and skeletal overgrows, indicate that the flow interior was a true lava which suffered rapid quench crystallisation and high-temperature devitrification at the time of emplacement into the subaqueous environment.

An individual flow is characterised by a crystalline silicous body and crystal-tuff carapace (Plate 5.5. .). The crystal-tuff is composed of angular to sub-angular feldspar phenocrysts and microphenocrysts, set in a fine grained tuffaceous groundmass. The top of individual flows merge with the base of the succeeding flow to give a bedded appearance at outcrop (Plate 5.5. .). At several localities one can observe the crystal-tuff to be flamed into the base of a coherent body. The lack of intervening sediment, or reworking of the crystal tuff carapace, suggests frequent eruption and emplacement, possibly one after the other with little time lapse between successive flows.

Interpretation

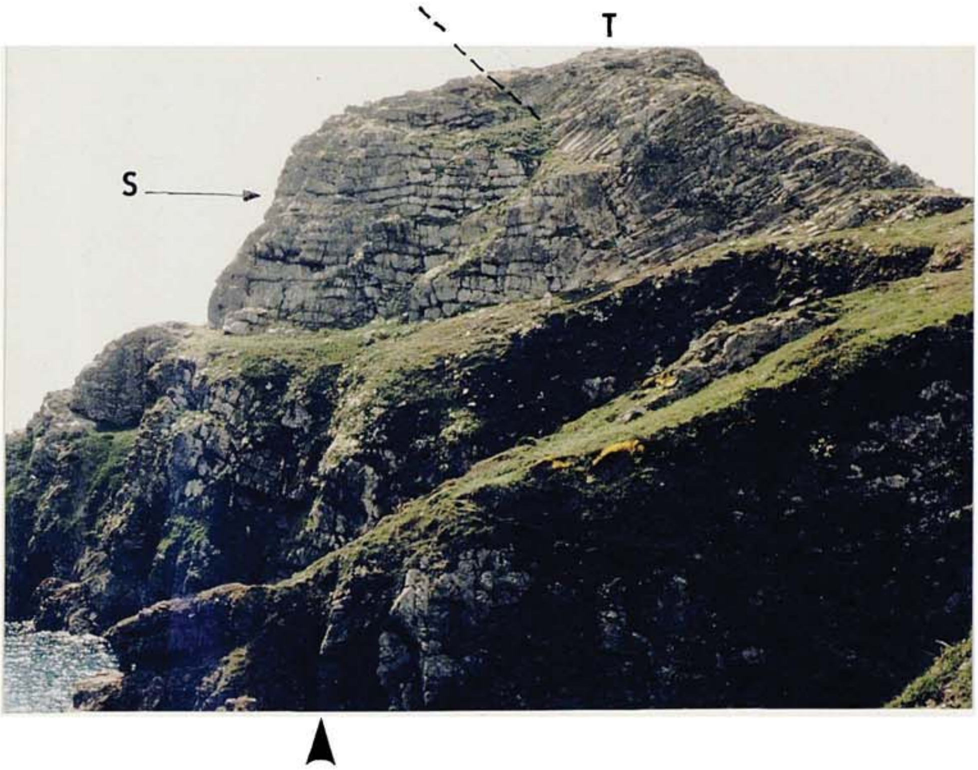
The apparent fluidity of the flows may relate to the high-hydrostatic pressures (500m-2000m water column) envisaged for the development of the rhyodacite pillow of the Porth Maenmelyn Volcanic Formation (Bevins & Roach 1979b). It thought that the flows would have been erupted at a comparable depth, which may have been greater than the PCL (*c.f.* Fisher 1984) for this specific style of silicic volcanism (the PCL for silicic lavas is dependant on many variables, see Fisher 1984). If correct, such a depth (pressure) would have had a dual purpose: firstly, it curtailed large-scale phreatomagmatic disruption inhibiting the expansion of volatiles, and secondly, retained the volatiles within the flow unit which maintained viscosity of the flow and so maintained mobility (Cas 1978). Why the flows suffered little quench fragmentation is uncertain. Recently, Howells *et al.* (1985) in assessing the emplacement mechanism of subaqueous ash-flow tuffs from the Capel Curig Volcanic Formation, N.Wales, suggest that the tuffs possibly maintained coherence as a result of the insulating effect of the Leidensfrost phenomenon (c.f. Mills 1984). Such an effect may have been applicable to the sheet-flow lavas within the Porth Maenmelyn Formation, inhibiting the effects of quench fragmentation therefore sustaining mobility. It is thought that the crystal tuff carapace reflects the elutrition of a fine fraction that was placed into suspension as an overlying cloud as the lava flow preceded.

5.4.2.2. INTRUSIVES

Evidence for contemporaneous intrusive activity associated with subaqueous volcanism is widespread throughout the Fishguard Volcanic Complex. Compositionally, intrusions have been demonstrated to range from basic to intermediate, with the development of gabbros and tonalites, and finer grained equivalents to the west of Fishguard (Bevins 1979, 1982). To the east of Fishguard, Lowman (1977) and Lowman & Bloxam (1981) document a suite of

Plate 5.5. Panoramic view of the field relationships and lithologies within the Porth Maenmelyn Volcanic Formation. This plate shows the development of sheet flow silicic lavas on the headland of Dinas Mawr (S) which are flat lying. The sequence is cross-cut by a tonalite dyke (T) the columnar jointing of which is coincidently approximates the thickness of individual sheet flow lavas. The entire sequence in thrust over vertically dipping basaltic pillow breccias, bedded silicic tuffs, and basaltic hyaloclastites (arrowed). The sequence is cross-faulted in the small gully in the center field of view (see Map 5).

Plate 5.6. Close-up view of a single sheet flow lava silicic lava on the headland of Yns Ddyd. The lava flow has two main parts: a coarse to fine crystal tuffs at the top which merges into the base of the overlying flow (recognised by its deep weathering profile) and a main flow body which is composed of quench textured silicic lava. At two localities on the headland (Plate 5.5) underlying tuffs were seen to be injected in small flame structures into the base of the lava flow.


T
S

intrusives, ranging in composition from basic through to intermediate, including gabbros, dolerites, diorites and microgranites. Whilst it can be demonstrated that the intrusive suite was contemporaneous with the development of the Fishguard Volcanic Complex (see below). It would appear also, that intrusive activity ceased in the whole region during late Upper Llanvirn times. This is indicated by the lack of field evidence for intrusions cross-cutting late Upper Llanvirn and Lower Llandeilo sediments of the Lower Town and Dyffryn Formations {4.6.}, and the Upper Llandeilo *N. gracilis* Shales east of Fishguard (Lowman 1977).

Within the Fishguard Volcanic Group two distinct intrusive 'types' are observed: those which can be inferred to have been emplaced at high-levels into the volcanic pile and associated sediments, showing evidence for interaction with poorly lithified sediments (see Kokelaar 1982), and larger coherent gabbro and tonalitic bodies (Map 1-4.). Whilst both intrusive types were coeval with the major phases of volcanism and generically related compositionally (Bevins 1982), the latter may possibly represent part of the more regional intrusive episode throughout SW Dyfed, reflecting regional extension.

Unlithified sediment – magma interaction: Throughout the coastal districts between Strumble Head and Maen Jaspis, one can observe the interaction between magma which had been emplaced at shallow levels into poorly consolidated sediment. The style of interaction is variable from bulbous sills and dykes, through to chaotic mudstone - magma mixtures (Lorenz 1984). There is overlap between both types (and the local development of pépérites), although they may be regarded as the 'end-member' products of the magma – wet/sediment associations (see Kokelaar 1982 for details). Such intrusives differ from the larger coherent intrusive bodies, in that they show signs of interaction with poorly lithified sediment, are generally smaller, aphyric to weakly porphyritic, and are highly altered. The latter may suggest some degree of 'spilitisation'.

Examples of *small bulbous sill* (and more rarely dykes) emplaced into poorly lithified sediments are common throughout the upper parts of Strumble Head Volcanic Formation. Such intrusives may resemble pillow lavas in that the intruded magma maintains an overall coherence in intrusive form. The sills show little evidence for hyaloclastisation or brecciation. Some of the best examples of this intrusive style can be seen at Aber Gwladus. Here, pillowed and bulbous basaltic sills are intruded into fine grained pelitic and porcellanitic mudstones, which underlie part of a basaltic bomb-lapilli-ash breccias (Plate 5.12.). The sills show little evidence for hyaloclastisation terminating in bulbous pillowed protrusion. The host sediment is strongly silicified, although lamination within the host sediment is only destroyed within a few centmetres of the magma-sediment interface. Such relationships are typical of the complex interrelationships of high-level sills emplaced into poorly lithified sediment, where preservation of the sediment and the intrusive form are maintained by the process of fluidization (*c.f.* Kokelaar 1982)

An excellent example of a *chaotically mixed mud-magma sill* is seen within the upper parts of the Strumble Head Volcanic Formation between Porth Sychan and Panty-y-Dwr. The sill can be observed in the cliff exposure at Pen-Capel Degan [S.M. 9080 4100] to the west, passing eastwards to Aber Clawdd Pridd [S.M. 9175 4080], Carn Halen [S.M. 9225 4060],, Panty-y-Bara [S.M. 9340 4055] and Panty-y-Dwr [S.M. 9345 4030], underlying the larger coherent dolerite which is depicted on Map (2). The most representative and accessible exposures are those at Pen-Capel Degan and Panty-y-Bara. The sill is fine grained and weakly porphyritic, variable in thickness from *c.* 1-3m, although highly conspicuous due to its involvement with argillic sediments. The true longitudinal dimensions of the sill are uncertain due to cross-faulting and discontinuous exposure, although at outcrop it is clearly seen at all the above localities, suggesting a strike length of ≈ 2.5km is appropriate. The intrusion is singularly unusual in this sense alone, comparable with the mixed mud-magma succession of the Dunnage Mélange, Newfoundland (Lorenz 1984). The nature of the sill is seen to vary along its strike length. It may locally show evidence for hyaloclastisation and auto-brecciation with the formation of pépérite and a hyaloclastite/sediment mixture with only the remnants of a sill (i.e. lobate protrusion) existing. Generally, however, it exhibits complex interdigitating contortions and folding which remains coherent in form with only minor hyaloclastisation along the intrusive/sediment interface. Such relationships, which at first sight appear incompatible, are now widely believed to owe their existence to fluidization of the host sediment and its insulating effect on the magma so as to inhibit direct chilling (see Kokelaar 1982).

Coherent intrusives: ***(Gabbros, Dolerites, Tonalites and Microtonalites);***

Throughout the Fishguard Volcanic Group, a variety of coherent intrusives bodies crop out within and underlying the extrusive successions and sediments which spatially underlie the Complex. Bevins (1979), Bevins & Roach (1982), Bevins (1982) and Kokelaar *et al.* (1984b), document the suite as gabbros/dolerites (basic), tonalites/microtonalites (intermediate); this terminology has been followed in the field (Map 1-6). Distinction between grain-size variants has been arbit arily distinguished. There appears to be little evidence at outcrop for internal differentiation of the intrusives, such as the norites, quartz norites, gabbros, and enstatites as described by Roach (1969) from the St. Davids – Carn Lliddi intrusion to the west. Cross-cutting relationships have been observed at several localities (Plate 5.7.). In all cases intermediate intrusives cross-cut basic intrusives, at no point has the present author seen evidence for the reverse, this may suggest compositionally discrete magmas at this level within the Volcanic Complex.

Tonalites and Microtonalites: Tonalites and microtonalites occur throughout the area around Porth Maenmelyn and Penbwchdy District (Maps 1 & 3.). Both lithologies are characterised by a feldspar-phyric mineralogy with minor quartz, no primary mafic's have been observed in thin-section, although sparodic areas of equant chlorite possibly represent mafic pseudomorphs. Petrographically, microtonalites contain abundant areas of interstitial mesostasis, whilst crystalline phases are dominated by fine seriate feldspar microphenocrysts. Dendritic and skeletal overgrowths attest to rapid cooling. Coarser grained tonalites (e.g. Penbwchdy Headland) are dominated by subhedral granular prismatic feldspar, reflecting the relatively prolonged cooling histories in comparison to their finer grained counterparts. K-feldspar have been identified from tonalite dykes, and may represnt secondary hydrothermal alteration, common to subaqueous silicic lavas and intrusives (Munhá *et al.* 1980). Prehnite, chlorite, stilpnomelane, white-mica and frambiodal Fe-rich sphene are developed, resulting from low-grade regional metamorphism. Fe-rich Ca–Al silicates are rare.

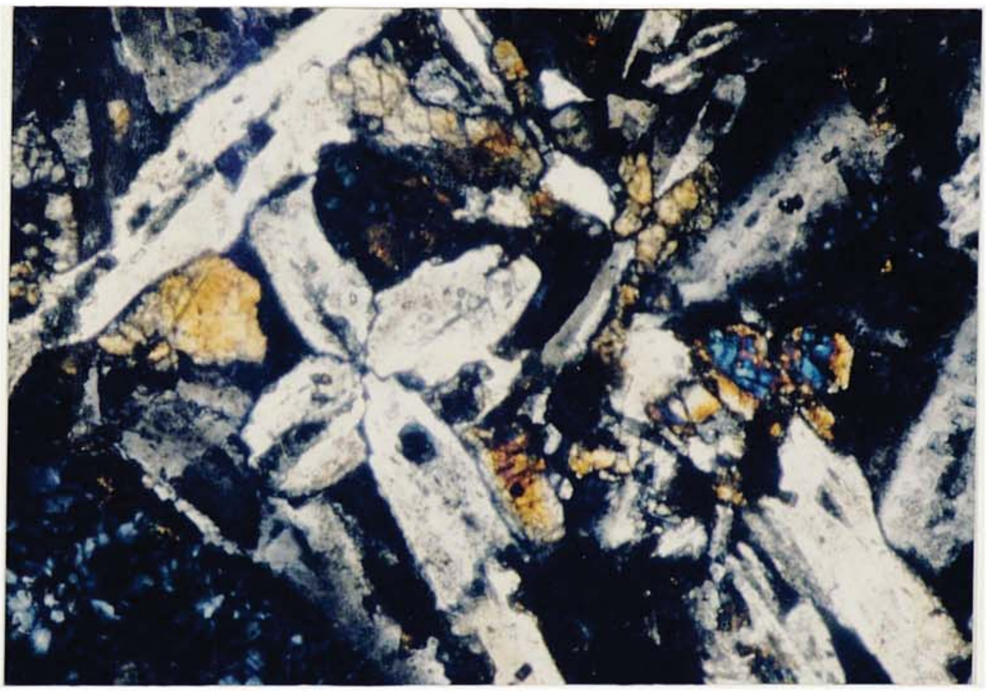
Gabbros and Dolerites; Dolerites occur throughout the upper parts of the Strumble Head Volcanic Formation (Map. 1). The largest forms a laterally continuous and concordant sill, traceable from the headland of Pen Caer eastwards to Carnfathach [S.M. 9380 4030] and inland to Pen-rhiw [S.M. 9410 3500], a collective strike length greater than 5km. Gabbros spatially underlying the extrusive successions on the Pen Caer peninsular (Map 1.). It is envisaged that such bodies represent the deeper part of the Complex, possibly representing high level chambers which feed hypabyssal dolerites. The mineralogy of basic intrusives is dominated by feldspar, clinopyroxene (augitic composition), and ilmenite- magnetite opaques (ore), apatite is a common minor accessory in coarser grained lithologies and pegmatitic areas. Kaersutite is locally present as poikilitic inclusions in clinopyroxene. Areas of interstitial mesostasis are recognised by equant voids infilled with chlorite, although presumably represented areas of glassy residue. Individual plagioclase and clinopyroxene crystals are generally equant, euhedral to subhedral and commonly equigranular, although seriate textures are common to dolerites. Sub-ophitic intergrowths of plagioclase and clinopyroxene, and more rarely poikilitic textures predominate. Intersertal textures are common within the gabbros. Feldspars are generally albitic in composition, although composition is assumed to be related to low-grade regional metamorphism, the precursor feldspar composition likely to have been of labradoritic composition (Bevins 1982). A range of low grade secondary minerals have developed in response to regional metamorphism (see Chapter 6).

Prior to the effects of regional metamorphism it is thought that the gabbros may have suffered

Plate 5.7. Cross-cutting relationships of tonalite (T) and gabbroic (G) intrusions within the Fishguard Volcanic Group at Garn Fawr [S.M. 8955 3875]. The relationship is of the tonalite cross-cutting the gabbro, as shown by the presence of abundance of small vesicules within the tonalite near the contact (above V) and a fine chilled margin to the tonalite which is only recognised in the feild. At all localities were cross-cutting relationships have been are observed tonalite (intermediate) cross-cuts gabbro (basic).

Plate 5.8. Photomicrograph of holocrystalline core from a basaltic pillow lava, showing typical intergranular textures between feldspar and clinopyroxene (Field of view 2.7mm)


C
T
V



low-pressure amphibolitisation and epidotisation by reactions with post-magmatic deuteric and/or hydrothermal fluids. Amphibolitisation is suggested by the common occurrence of secondary hornblende+quartz+apatite assemblage seen locally (in thin section) in areas of mesostasis and replacing clinopyroxene. Epidotisation is indicated by the development of epidote metadomains (*c.f.* Smith 1968a) which can be clearly observed in the field as a lime-green waxy replacement, common, although not exclusive to pegmatitic areas. Both styles of alteration are not compatible with regional metamorphism (see Chapter 6).

Felsic microdykes: Thin felsic dykes varying in thickness from 5mm to 5cm are seen to cross-cut (with a brittle fracture) the gabbros between Carn Gelli and Llwanda; they are particularly numerous in the area Henner cross [S.M. 9300 3810]. The microdykes are microporhyritic suggesting a relatively evolved magma. Fine chilled margins are present. Associated with the microdykes, is the development of a suite of secondary amphiboles. Two species predominate sodic-calcic amphiboles of the winchite-richterite series and ferro actinolitic-hornblende. It is thought that these amphiboles may reflect alteration by Na-rich fluids associated with dyke emplacement.

5.4.2.3. Basaltic volcanism and intra-volcanic Complex sedimentation.

Basaltic volcanism within the Fishguard Volcanic Group (i.e. Strumble Head Volcanic Formation) was characterised in the main by effusive pillowed basaltic flows, although pillow breccias, sheet-flows, and volcaniclastics (i.e. Carreg Gybi Member) developed locally. The deposits represent an interplay of extrusive and sedimentary processes within the subaqueous environment. This following section outlines the stratigraphy of the sequence in terms of the Strumble Head Volcanic Formation and the Carreg Gybi Member (Figure 5.1., Map 2 & 5) and then goes onto discuss several aspects of the pillow lavas on the Pen Caer Headland. Other lithologies are noted.

STRATIGRAPHY

A) Strumble Head Volcanic Formation (SHVF).

The Strumble Head Volcanic Formation occupies the majority of the area on the Pen Caer Peninsular (Maps 1-2). The Formation is composed of a thick sequences of pillow lavas, high level dolerites and subordinate hydroclastics of varying affinity. Inland exposure on the Pen Caer Peninsular is exte emly poor, although coastal exposures between Porth Maenmelyn - Strumble Head (perpendicular to strike) and Strumble Head to Maen Jaspis (parallel to regioanl strike) allow an excellent insight into the various products of subaqueous basaltic volcanism. The thickness of the Strumble Head Volcanic Formation has been widely regarded by previous workers as approaching 1500m. This thickness is the Formation's approximate development between Porth Maenmelyn to Strumble Head. It is tempting to suggest that a thickness far less than 1500m may be appropriate, in some much as, the most complete succession unaffected by thrusts, occurs between Pwll Arian and March Bach, a distance no greater than 350m. However, the homogenous character of the Formation and selective deformation of intra-formational sediments does not allow the identification of repetition, as can be demonstrated elsewhere (Maps 5 & 6). Cross faulting also effects the distribution of the Formation.

The inland exposure of the Strumble Head Volcanic Formation, south of the Pen Caer Peninsular is extremely poor. Pillow lavas, basaltic tuffs and associated gabbros and dolerites are locally evident. However, boundaries are difficult to define, whilst the relationship with other successions is largely unknown. It is suspected, that the observed thinning of the Formation southwards from Pen Caer is primarily the result of Caledonide tectonism. However, tectonism may have excentuated an original volcanotectonic depression, in so much as, Kokelaar *et al.* (1984a) visualise the basaltic lavas as having developed within a 'graben-like' structure, within which subsidence kept pace with the extrusion of magma. The presence of interbedded pelitic/porcellanitic mudstones and various volcaniclastics which were deposited in the main by turbidity currents (see below), supports the contention that basaltic volcanics of the Strumble Head Volcanic Formation are unlikely to have formed any marked topographic feature on the sea-floor (Kokelaar *et al.* 1984a).

No basic volcanics are evident at south of Pen–rhiw, on the southerly limb of the Llwanda anticline, whilst there is no obvious feature to suggest their lateral continuity south-westwards towards St. Nicholas. It seems likely however, that extrusives of the Strumble Head Volcanic Formation continue for some distance, to a point where the entire successions is fault bounded against Lower Llandeilo sediments by the Goodwick fault (Map 1.). No continuity is offered by the basaltic volcanics south of Goodwick.

(b). Carreg Gybi Member (CGM). *Type section [S.M. 9030 4130], Thickness (≈ 30m).*

The Carreg Gybi Member represents a thick development of interbedded silicic volcanoclastic turbidites, pyroclastic debris flows, marine shales, and porcellanitic mudstones. The Member overlies the Strumble Head Volcanic Formation on the headland of Carreg Gybi [S.M. 9030 4130]. A repeated succession is seen along the cliffs at the west of Porthsycan [S.M. 9065 4085], much of the which is complicated by cross-faulting, although the nature of succession is almost identical to that exposed on the Carreg Gybi Headland. At its type locality, the base of the member is recognised by thin mudstones and basltic hyaloclastite which rest on pillow lavas of the Strumble Head Volcanic Formation.

DESCRIPTION OF LITHOLOGIES

Pillow Lavas

Pillow lavas and tube networks constitute approximately 70% of the exposed seqence. Following recent observations from DSDP cores and actively forming pillow lavas on the mid-Oceanic ridge and young seamounts (i.e. Moore *et al.* 1971, Moore 1975, Ballard & Moore 1977, Fornari *et al.* 1979, Lonsdale & Batiza 1980), with comparative studies on ancient successions (e.g. Dimroth *et al.* 1978, Wells *et al.* 1979, Yamagishi, 1985), it is now widely recognised that pillow lavas form by a process of 'budding' from inter-connected tube networks. Early interpretations of discrete sac-like bodies, occur only in rare circumstances. The nature and observed geometries of the pillow lavas in the Strumble Head Volcanic Formation accord well with such a processes (Plate 5.9 ., 5.10.)

Morphology and surface textures: In two dimensional cross-section individual pillow structures are readily identifiable, as a pattern of surfaces delimited by closed curves, either in contact or separated by a matrix of variable thickness, and often composition. The two dimensional geometric shape is variable, although all essentially possess a characteristic elliptical to sub-rounded convexo-concavo morphology and cuspate bottom, with sufficient irregularity to maintain a close fit. Shape varies in accordance with the most commonly documented discriptive geometeries (e.g. bun-shaped, bean-shaped; see Vaught 1975, Dimroth *et al.* 1978). It would appear that shape is governed by several factors (i.e. proximal/distal domain flows), although not least, by the underlying surface topography, with pillow tubes molding themselves on pre-existing lower surface morphologies, draping and sagging over surfaces whilst still in a semi-plastic state. In cross-section, pillows internally exhibit characteristic features such as hyaloclastic margins, radial and locally subconcentric jointing, vesicules, and in larger pillows, zoned crystal growth. Hollow or 'drained' pillows, as described in both modern and ancient pillow lava successions (e.g. Ballard & Moore 1977, Grenne & Roberts 1983) are generally rare. This may suggest high extrusion rates, or insuffïcent topographic relief so as not to allow pillows to drain away, although it is thought more likley to reflect the low preservation potential of such lithologies.

The diameter of individual pillow lavas is seen to vary from as little as 10cm to greater than 4m; reflecting variation in extrusion rate from the site of eruption, distance from the sight of eruption, and relative position within a tube network. Three dimensional exposures of the pillow lavas accord well with documented pillow tube-networks. It is generally accepted that molten lava is conveyed from the site of eruption by lava tubes and interdigitating sinuous tube networks, which fracture along spreading and tensional cracks. Such a process readily accounts for relationships shown in Plate 5.10

Primary surface textures and structures are commonly preserved. Many unweathered surfaces have a weak corrugated texture (Plate 5.10) comparable to the *'ropy winkles'* of Yamaghisi (1985). A variety of spreading, tensional, and contractional cracks are developed. Contractional cracks develop once growth has stopped, whilst others are formed during growth of the pillow flow (Yamaghisi 1983). Tensional cracks are commonly observed on the sides of larger pillow tubes (Plate 5.10). It is likely, that during pulsating supplies of fresh lava through a single master tube, that pre-existing tensional cracks may be areas of weakness and allow small lateral buds to develop (Plate 5.10). Contractional fractures are developed in the majority of pillow structures, the most common of which are those generally radial to the pillow cross-section restricted to the outermost part of the pillow. Locally sub-concentric ring fractures may also be observed. The relationship of both fracture sets forming conjugate patterns appear to have a significant control in the development of pillow-block breccias and pillow front/foot breccias (see below).

Pillow Lava Petrography: The chemistry of individual primary phases within basaltic pillow lavas from the Fishguard Volcanic Complex has been documented in detail by Bevins (1979, 1982). The primary igneous mineralogy is essentially clinopyroxene (augitic composition, {6.4.1.a.}), plagioclase feldspar, and ore (opaques), with secondary chlorite assumed in part to be after a sideromelane precursor. A wide variety of secondary minerals have developed in response to low-grade regional metamorphism (see Chapter 6). Prior to the affects of such low-grade metamorphism, the pillow lavas are likely to have suffered alteration due to a high sideromelane content. It being probable that the extrusive basic lithologies suffered palagonitization at an earlier stage, possibly followed by the incipent development of zeolites.

The radial geometry of pillow lavas allows textural zonation which can be attributed to a range of cooling histories. Whilst cooling rates are the major controlling factor in textural development, pre-eruptive history, confining pressures prior to eruption controlling vesicularity, temperature of eruption, melt bulk composition, volatile content, and oxygen fugacity, may also be influential (Schiffman & Lofgren 1982). Three main textural zones can be discerned: a) hyalocrystalline rim, b) hypocrystalline spherulitic and dendritic zone characterised by skeletal growth, c) holocrystalline core. The hyalocrystalline and hypocrystalline zones are present to all pillows whilst the holocrystalline core is generally restricted to larger pillows. The general textural features can be summarised as follows :

hyalocrystalline rim: The thickness of the hyalocrystalline rim varies from 1-4cm, composed primarily of chlorite with accessory sphene and epidote. Chlorite commonly exhibits a colloform-radiate habit, possibly psuedomorphic with respect to a pre-existing perlite texture.

Vesicules are rare, and where observed are small and infilled with coarse radiate chlorite and calcite. Brittle subconcentric micro-fractures are locally developed, commonly infilled with calcite, chlorite and small amounts of K-feldspar. Rarely can such micro-fractures be observed to pass throughout the entire rim, although they may pass downwards and out into the hypocrystalline zone. Such micro-fractures are thought to represent a primary texture resulting directly from quenching and contraction of the once glassy margin.

Crystalline phases within the hyalocrystalline margin are rare. Towards the outer margin fine euhedral tabular plagioclase crystals (<1mm) are locally evident, although show no signs of dendritic or spherulitic overgrowths. Isolated spherulites can be observed locally. As one passes inwards towards the base of the hyalocrystalline rim, fine euhedral plagioclase crystals are present, although co-exist with larger (2-7mm) partially rounded and resorbed pre-eruptive plagioclase phenocrysts (*'xenocrysts'* of Bryan 1972). Such pre-eruptive phenocrysts are commonly shattered, although fragments are not dispersed and original crystal outlines are still discernible. This shattering of pre-eruptive phenocrysts within the hyalocrystalline rim is thought to represent a primary texture reflecting hydroclastic auto-brecciation as a result of 'thermal shock' related to quenching and rapid cooling at the lava-water interface.

At the base of the hyalocrystalline rim, a fine veneer (2-9mm) consisting of plagioclase and quartz in granular polygonal aggregates has been observed. This small veneer impinges with an ameboidal relationship upon the rim. The true textural relationships of such aggregates is invariably obscured by the development of secondary minerals, although it may represent a variolitic zone of recrystallized coalesced plagioclase spherulites (Furnes 1973). Its thickness, granular morphology and relationship is comparable, although spherulitic development is difficult to discern. Large pre-eruptive plagioclase phenocrysts (2-4mm) are conspicuous in their relative abundance, exhibiting shattering and resorption textures, suggesting disequilibria with the crystallising melt. A reason for the general abundance of plagioclase pre-eruptive phenocryst within the variolitic zone and hyalocrystalline rim, compared to their relative scarcity throughout the interior of pillow lavas, remains largely uncertain. Similar plagioclase crystal concentrations along the rim and centres of pillows have been observed within modern and ancient pillow tubes (Duffield 1969, Wells *et al.* 1979), attributed in part to flow differentiation and centrafugal flow by analogue to similar concentrations in dykes. A further explanation, is that the hyalocrystalline rim and variolitic veneer, have 'frozen' the pre-eruptive constituent mineralogy (i.e. magma mineralogy) of the lava flow within the glassy carapace at the time of eruption, prior to internal crystalisation.

hypocrystalline zone: Passing sharply inwards from the variolitic veneer, the hypocrystalline zone volumetrically dominates most of the pillow lava cross-section, although its thickness is dependant on the pillow diameter, which controls cooling rates and thickness of individual zones. Vesicules may be concentrated towards the outer margin of this zone commonly infilled with chlorite, calcite or quartz, either being monomineralic or admixture of these phases in varying proportions. Textural development of the infilling phases is commonly, colloform and coarse radiate growth, or polymineralic rhythmic colloform banding, although fine radiate and idiomorphic textures are observed. In addition to chlorite, calcite and quartz, colloform prehnite, prismatic and idiomorphic pumpellyite, granular and colloform sphene, prismatic epidote, and prismatic K-feldspar are observed. This zone is characterised by plagioclase microlites whose textural development is characteristic of rapid quenched crystallisation; exhibiting skeletal, spherulitic and hollow crystal growth, set within a matrix which is composed of fine grained chlorite and granular sphene, presumably after a sideromelane/palagonite precursor. The plagioclase microlites are locally seraite (on the scale of a thin section), although an overall increase in grain size and relative abundance towards the centre of the pillow is observed. Microlites typically exhibit variolitic divergent fan growth, bow-tie, sheath, and radial spherulitic growth, and occasionally fine hollow crystalline growth. Clinopyroxene is rarely developed, although this may be due to pervasive chloritisation and fine grain size (spherulitic fibres may have had clinopyroxene intergrowths). Microphenocrysts (>2mm) are generally tabular to skeletal, commonly exhibiting dendritic and swallow-tail overgrowths. Locally, glomerophyric clusters composed of skeletal and tabular plagioclase and subhedral clinopyroxene occur. The textural development of glomeracrysts are generally intersertal to sub-ophitic, similar to pre-eruptive phenocryst glomerophric clusters from mid-oceanic basalts (Kirkpatrick 1979a). Skeletal overgrowths on several plagioclase glomerocrysts suggest that at least some crystal development is post-eruptive, whilst limited evidence for resorption may indicate partial equilbria with the melt. Pre-eruptive phenocrysts of plagioclase whilst present, are rare, being distinguished by their rounded and embayed nature, and lack of spherulitic and dendritic overgrowths.

holocrystalline core: Passage from the hyalocrystalline zone to the holocrystalline core appears gradational, there being no definite boundary. The zone is characterised by an increase in grain size as one passes towards the pillow centre, although in pillow lavas with a small cross-sectional area it may not be developed. Clinopyroxene can be seen to be increasingly developed as anhedral aggregates interstitial between microlites and microphenocrysts. Towards the centre of the core, plagioclase is generally tabular and may locally show swallow-tailed overgrowths; skeletal crystals whilst present are generally rare. Clinopyroxene is relatively abundant, as anhedral plates. Intersertal, equigranular, and intergranular textures predominate, although locally, coarse spherulitic growths can be seen. Interstitial areas between crystals are commonly filled by chlorite, although opaques and fine plagioclase microlites may be seen, possibly representing late stage crystallisation of residual melt in areas of mesostasis.

The textures observed and their zonation through the pillow lava cross-section, can be attributed to a range of cooling histories. Prior to extrusion, it appears likely, that the pre-eruptive melt had developed to a point where plagioclase feldspar and minor clinopyroxene had started to crystallise. Upon extrusion onto the sea floor, the basic magma suffered rapid quenching at the lava-water interface followed by the almost instantaneous development of a hyalocrystalline carapace. Pre-eruptive crystals w ere 'frozen' in this carapice and suffered autoclastic brecciation as a result of thermal shock. Development of plagioclase microlite followed and small plagioclase microphenocryst crystallised towards the outer margins as cooling progressed The hypocrystalline zone developed possibly over a range in cooling history, although textures and crystalline forms reflect rapid crystallisation. The holocrystalline is the progession from the hypocrysalline zone reflecting the more static cooling conditions, with clinopyroxene developing in small amounts.

Vesicules and Scourious pillows: Jones (1969) used the pillow lavas on the Pen Caer Peninsular, upon which to test the hypothesis of an inverse relationship between vesicule content and depth of water. He concluded, that a positive correlation existed between the pillow lavas at Porth Maenmelyn (low vesicularity, depth of eruption ≈ 2km) and those that outcrop along the coast towards Strumble Head as one youngs up section (increasing vesicularity, depth of eruption ≈ 10's metres). Whilst not vigorously investigated, such a correlation was not observed during this study, whilst the majority of pillow lavas around Strumble Head and Carreg Gybi are low vesicularity pillows. Vesicules occur in varying degrees in all pillow lavas. Distribution of vesicules within the pillow cross-section may show a slight concentration towards the outer surfaces. In most cases, vesicules volumetrically make up less than 5% of pillows, although locally they may constitute more than >20% forming scoriaceous pillows. The majority of vesicules observed (in thin-section) are spherical to subrounded in shape and small in cross-sectional area (invariably < 5mm, and generally < 3mm), a feature typical of deep marine pillows (Wells *et al.* 1979). Such observations would suggest that eruption was significantly below the PCL for basaltic volcanism (≈ 200m, see Fisher 1984). However, scouriaceous pillows remain largley enigmatic, in the sense that they are likely to represent anomolous volatiles concentrations significantly above the 'background', as observed by the vesicular content of the majority pillow lavas. A possible explanation for their origin is that it is increasingly accepted that volatile gradients exist within magma chambers (for details see; Fisher & Schmincke 1984) and it may possibly be, that scouriaceous pillowed flows reflect variable magmatic volatile partial pressures. Such a suggestion is based on the inabilty to visualise, interbedded scourious and weakly vesicular pillowed flow units [S.M. 8820 3980]

Plate 5.10. Two dementional exposure of pillow lavas from the Strumble Head Volcanic Formation at Ogof Phillip. Note thin hyaloclastic margin and chilled rims, below which vesicule concentrations can just be discerned. Note how shape is dictated in the main by the underlying pillow

Plate 5.11. Three dementional exposures of pillow lavas from the Strumble Head Volcanic Formation at Trwyn Llwyd. The plate shows the development of a short lateral bud from a master tube. The corregated surface textures are comparable to the 'ropy wrinkles' of Yamighisi (1985). Arrow at base of plate is showing a typical tensional crack.

as arising from rapidly flucuating hydrostatic load pressures, which would produce a similar effect.

Inter-pillow matrix (Triple-Junction Fill): The matrix that occupies the 'triple- junctions' between pillow lavas is variable in habit, composition, volumetric abundance, and distribution. Four distinct matrix types are evident within the Strumble Head Volcanic Formation, of which more than one type may cohabit the triple-junction between individual pillows, they are: hyaloclastite matrix, chert - silicous matrix, pre-eruptive sediment matrix, post-eruptive sediment matrix. For convenience, although not strictly regarded as a matrix lithology, intra-flow 'rip-up' clasts (cognate lithics) are also noted.

Hyaloclastite matrix; Hyaloclastite matrix (Rittaman 1962, Honnorez & Kirst 1975) is the most common of matrix type, being typical of most documented subaqueous basaltic flows. The development of the matrix is variable in thickness, from thin selvedges which may fill triple-junctions, to a matrix that is sufficently thick so as to give the associated pillows an isolated appearance in cross-section; the latter is rarely developed and usually confined to small lateral buds. It is generally accepted, that such a matrix represents the spalling of hyalocrystalline shards from the glassy carapice during development of the lava flow. However, only rarely can shardic morphologies be discerned, the effects of low-grade metamorphism obscuring much of the primary fabric.

Chert matrix; Inter-pillow chert (non-organic) is common throughout the pillow lava successions. Its volumetric abundance may locally exceed the hyaloclastite matrix, although its distribution appears random. It exhibits a variety of colour from black to dull grey (porcellanite), apple green (chryosphase), red to dull brown (jasper), the likely consequence of variable impurities and oxidation states of opaques. Colour is however, haphazardly distributed, two or more coloured varieties may be present within the same triple-junction.

The chert commonly replaces pillow margins and hyaloclastite matrix, and more rarely replaces pillow interiors along radial fractures. All cherts show varying degrees of recrystallisation (in thin-section). The source of the silica which precipitated the chert is likely to be hydrothermal in origin, sourced from silica-rich fluids associated with extrusion of the lava on to the sea-floor and/or fumaroles, situated along structural weaknesses which vented silica rich fluids into the volcanic pile. Evidence for the latter may occur at Ogof Philip, where a discrete chertified band $\approx$ 4-5m runs vertically N–S across the pillow lavas in this region, replacing large areas of the host rock in a discrete zone. It is envisaged, that such areas may have represented structural weakness within the sea-floor.

Pre-eruptive sedimentary matrix; Where individual flows can be delimited by inter-flow sediments, it is common to observe the flaming of sediment into the overlying lava, filling the interstices between individual tubes and lobes. Whilst this flaming is in general confined to the lowermost parts of individual flow unit, pre-eruptive sediment has been observed to occur up to 3m above the basal lava/sediment interface within a flow at the south end of Carreg Onion Bay.

It is thought that this may reflect, either loading due to density contrasts, or fluidization (*c.f.* Kokelaar 1982) of wet sediment into the flow interior subsequent to emplacement of the lava on the sea-floor. It would be realistic to assume that both processes were operative. Gravitational loading of pillow tubes into poorly consolidated sediment would undoubtedly have occurred, whilst m etre-scale flaming of sediment into pillow interiors suggests that the latter processes of fluidization may have been locally operative.

Post-eruptive sedimentary matrix: Sediments occupying the upper flow surfaces of individual flows are locally preserved, particularly within larger bulbous pillowed flow surfaces; although weathering of surfaces tends to inhibit good preservation. Within the pillow lava successions around Dritwig and Penrhyn, examples of post-eruptive sedimentary fill of pillowed surfaces can be observed. Such sediment-fill can give an insight into the processes which were operative during interflow periods of short duration, rather than prolonged periods of quiescence from active volcanism where the presence of marine shales reflect a return to the regional background conditions of sedimentation.

At Dritwig the pillows are mega bulbous pillows (> 3 metres in diameter) which create large large conical triple junctions on the upper sufaces, tapering into the flow interior to depths greater than 2m. The sediments which fill this relatively large space are fine to medium bedded medium to coarse grained parrallel laminated tuff, composed of shardic debris rare feldspar crystals. In thin section, extensive chloritisation and prehnitisation, obscures much of the original sedimentary fabric. Lamination occurs throughtout the fill. Weak grading of individual beds and equant detrital mineralogy points to a possible origin as near vent 'subaqueous ash-fall', where material was placed into suspension, probably generated from turbulent clouds above eruptive fissures. Small scale soft-sediment deformation is common to such horizons, readily recognised by small conguate fracture systems with concordant drapes. It would appear likely that such deformation reflects seismic activity which would undoubtedly have occurred frequently during eruption and development of the volcanic complex.

Intra-flow rip-up clasts: Locally within the pillow lavas on the Pen Caer headland, one can observe lithic blocks which have been incorporated within pillowed flows. The best successions where intra-flow rip-up clasts are observed, crop out on the headland and wave-cut platform of Pen Capel Deggan [S.M. 9080 4080]. Here, a thick pillowed unit, contains lithic blocks up to 1m in length which rest between tubes. The blocks are composed of fine grained parallel laminated porcellanitic muds, equant in shape, commonly tabular. Their shape suggesting either minimal transport distances or that the blocks where carried as 'suspended' lithics within the advancing flow .

Whilst such lithic blocks, can readily be interpreted in terms of accidental clasts incorporated, probably by erosion at the flow head, or possibly from sediments within the immediate vicinity of the eruptive fissure, their equant shape and size suggests that the sediments were lithified, or partially lithified, in order to maintain clast angularity and coherence. This may indicate rapid or accelerated lithification of the sediments on the sea-floor. However, this may not be unexpected, as Kastner & Siever (1983) demonstrate that siliceous and pelitic turbidic sediments which have been intruded by high-level basaltic sills within the Guaymas Basin spreading centre, Gulf of California, suffered diagenetic changes over thousands of years which otherwise would normally take millions to tens of millions of years.

Non-pillowed Flows

Pillowing of extruded basaltic magma whilst by far the most common lava form, was not ubiquitous. During development of the Strumble Head Volcanic Formation, non-pillowed flows developed locally in a sheet like manner. Whilst such flow-types are not numerous, they are locally developed around the headland of Pen Caer (e.g. the columnar spilite of Thomas & Thomas 1956) and along the coastline towards Pen Bush.

Thickness of non-pillowed flows is variable from several metres to 10's of metres. Within thicker flows crude, columnar jointing may locally be developed. They can be distinguished from high level dolerite sills, by their finer grain size, vesiculation, flat undulatory to weakly pillowed upper surface, thickly chilled contact surfaces, and buff green colouration (i.e. high original sideromelane content now chlorite). Locally small flow-top and flow-foot breccias are seen [S.M. 8840 4060; Carn Melyn]. It has been suggested that non-pillowed, or sheet-flows, represent high extrusion rates, where the high eruption rate inhibits the formation of crusts that are strong enough to contain fluid lava within pillow structures (Grenne & Roberts 1983). Such rates of extrusion are most likely to occur close to the feeding fissures, and conversely such flow types have been regarded as 'proximal' domain flows (Dimroth *et al.* 1979) which grade out into pillow facies, which themselves gradually decrease in size. At present no such sequence has been identified in the Fishguard Volcanic Group.

Basaltic pillow breccias and other hyaloclastic deposits

Pillow breccias (*c.f.* Carlisle 1963) occur through out the Fishguard Volcanic Complex, being confined in general to the basaltic successions on the Pen Caer peninsular. Bevins & Roach (1979) have discussed 'isolated pillow breccias' in association with pillowed rhyodacites at Porth Maenmelyn. Here a brief description is limited here to those of basaltic affinity. The best developed and accessible pillow breccia sequences crop out around Pwll Arian [S.M. 8050 4040], Porth Sychan [S.M. 9285 9045], Aber Gladyws [S.M. 9240 4150] and Carregwastad Point [S.M. 9265 4160] (Map 3 & 6). From these localities several pillow breccia types can been identified. They are briefly considered here, under the headings of: pillow fragment breccia, isolated sediment pillow breccia (pépérites and mass-flows); flow-front and flow-foot breccia, and bomb-lapilli-ash breccia.

Pillow fragment breccia (Staudigel & Schmincke 1984). The best examples of a pillow fragment breccias occurs at Pwll Arian [S.M. 8050 4040]. The breccia occurs as a thick tabular unsorted units, associated with mudstones, true pillowed flows, and high-level doleritic intrusives (Map 2.) The breccia is composed of lapilli to block sized basaltic clasts, many of which can be recognised as scoriaceous pillow fragments. The matrix is generally than 15% of the rock type, commonly giving clasts an isolated appearance. The matrix is comprised of fine grained hyaloclastite, although primary shardic textures are obscured by the effects of low-grade metamorphism. Vesicules can be observed in parts of the fragmental LITHOLOGY, AND may be better termed hyalotuff (*c.f.* Honnorez & Kirst 1975). Lithologically similar breccias have been described as 'broken pillow breccias' (Carlisle 1963), 'pillow fragment breccias' (Fisher & Schmincke 1984), and 'pillow breccias' by others (e.g. Barret & Spooner 1977), and their commonly explained as the 'mechanical unrimming' of pillow lavas (Carlisle 1963). However, the high percentage of hyaloclastite and scoriaceous pillow fragments, may suggest that fragmentation was also by hydroclastic processes.

Isolated sediment breccia (pépérites and mass-flow); Two breccias associated with sediments are seen within the Strumble Head Volcanic Formation. Pépérite pillow breccia (*c.f.* Schmincke 1967) can be observed on east side of Porth Sychan, were large bulbous pillow have loaded (possibly aided by fluidization (*c.f.* Kokelaar 1982) into fine argillic sediment. The base of the bulbous pillowed flow is fragmented to form a chaotic intermix of, pillow fragments, basaltic hyaloclastite, and mudstone, comparable to the 'pépérite breccia and fused tuffs' described by Schmincke (1967) in subaerial basaltic flows. A further basaltic/sedimentary breccia is seen on the headland of Trwyn Llwyn to the east of Pwll Lloug. At this locality basaltic fragments, some of which resemble pillow fragments, sit 'isolated' within a structureless mudstone matrix. Basaltic fragments vary in size from block to lapilli, and are subangular. No sorting is apparent, whilst there is little evidence for the hyaloclastisation of the basaltic lava fragments. The lack of evidence for structure within the host sediment and evidence for hyaloclastisation suggest that the mixture was deposited as a sedimentary gravity flow, comparable to the mass-flow, sediment supported mafic and ultramafic breccias described by Gianelli & Principi (1974).

Plate 5.12. Bomb/Lapilli/Ash Breccia at Aber Gwladus. The breccia unit is approximately 9m thick. The base of breccia rests above bulbous basaltic sills (B) which are intruded into mudstones. Note the discordance between a crude bedding in the breccia and the sill/sediment mixture. Further note, the decrease in lapilli and block size pyroclasts from right to left, and splatter bombs in top left hand corner. The breccia is believed to reflect the development of a small tuff-cone on the sea bed.

Plate 5.11. Pillow front-breccia from the headland at Carregwasted Point. The hammer is sitting on the breccia at the front of a pillowed flow network. Note conjugate fracture patterns on the pillow tube (T) to the right of the breccia It is thought that flow front breccias may owe much to the development of such fracture patterns.

B

T

Flow-front and Flow-foot pillow breccia; As the names imply, these breccias are observed at the base and front of individual flows. Such breccias occurs sporadically throughout the Strumble Head Volcanic Formation and are thought analogous to the 'flow-foot rubble breccias' of Ballard & Moore (1977) and Grenne & Roberts (1983). Examples of both breccia types can be seen on the headland of Carregwasted Point.

Flow-foot pillow breccias vary in thickness. They generally of the order of several ten's of centimetres, although locally they may be thicker. The breccia, is invariably composed of equant lapilli sized clasts, forming a basal brecciated veneer to an overlying flow. The process by which such a breccia may from, can be visualised as the accumulation of pillow talus at the front of the flow unit, possibly generated by the avalanching of broken pillow fragments and isolated budded pillows which became detached from the tube network, this accumulation subsequently becomes over-ridden as the flow-front progressively advanced, leaving a basal 'lag-like' deposit at the foot of the flow.

Flow-front pillow breccias are similar in lithological character to the flow-foot breccias, being composed of equant lapilli to block sized clasts (Plate 5.11.). The formation of flow-front breccias can be visualised, as the mechanical brecciation of talus accumulation and flow-front tubes which are pushed by the advancing flow rather than over-ridden to produce a flow-foot breccias. Such a process has been termed the 'bull-dozer' effect, by Furnes & Fridleifisson (1979) in discussing pillow block breccias from Iceland.

Bomb-lapilli-ash breccia: This breccia type is the most complex observed during this study. The best example of this breccia type crops out in the cliff sections at Aber Gladyws (Plate 5. 12.). The breccias is comprised of vitroclastic fine to coarse ash and lapilli and block sized bombs. The breccia at Aber Gladyws forms a unit ≈ 9m thick, overlying bulbous sills which were intruded in poorly lithified marine shales. The breccia fines up in two dimensional outcrop, and is attributable to depletion in lapilli and block size pyroclasts. The top of the breccia is capped by drained and hollow pillow lavas suggesting that the unit formed a small topographic feature on the seafloor. A slope is also supported by the breccia unit having a weak fabric (analogous to crude bedding) which is at angular discordance with the lower pelitic sediments. The presence of bombs, the angular discordance, coupled with the gross fining-up (Plate 5.12) is interpreted to reflects development of breccia in manner similar to 'rim beds' which develop during construction of 'Tuff Rings' (see Fisher & Schmincke 1984).

The breccia at Aber Gwladus, is visualised as having developed in response to ejecta being emplaced from a lava-fountain. It is thought that the nature of the eruption producing the deposit may be comparable to a small 'Strombolian'–like event, or events, with pyroclasts being ejected and deposited whilst still hot, accumulating to form a small shallow tuff cone around a diatreme and source conduit, comparable to the lava fountain breccias documented from the Troodos Ophiolite by Schmincke *et al.* (1983).

Basic Hyaloclastites and Volcaniclastics: Locally within the Strumble Head Volcanic Formation, thin bedded basaltic volcaniclastic form discrete sedimentary horizons, which can traced over considerable distances. Some of the best examples observed during this study, underlie the Carreg Gybi Member on the headland of Pen Caer (Plate 5.13). Similar successions occur around Pant y Dwr. Bedding thickness is generally on the centimetre scale. The nature of bedding is variable, although they all commonly posses structures, such as small scoured bases and faint cross-lamination which suggests that they where deposited by gravity driven currents, rather than as ash-fall units.

The pillow breccias and basaltic hyaloclastites of the Strumble Head Volcanic Formation, suggest that variety of processes where operative during the development of the volcanic sequence. It is thought that small strombolian-like events may have formed small tuff cones on the seafloor, composed predominantly of juvenile pyroclastic eject. Current reworking of such deposits may given rise to the development of the turbidic volcaniclastics. Pillowed flows, locally produced pillow fragment breccias developing in response to *in situ* auto-brecciation producing lava fragmentation and thick accumulations of hyaloclastic debris by essentially hydroclastic processes. Flow movements developed autoclastic flow foot breccias at the base of flows, whilst locally flow-front breccias developed in response to avalanching and 'bull dozering' of pillows at the front of advancing flow. Talus accumulation of block breccias occurred at the base of small submarine slopes, whilst slope failure and slumping, possibly triggered by seismic activity, can be envisaged as producing small mass-flow submarine lahars, depositing an unsorted fine grained sediments and lapilli-block breccias mixtures.

Intra-Formational sedimentation (Carreg Gybi Member)

Within the Strumble Head Volcanic Formation, discrete sedimentary horizons are scarce. This may reflect relatively high extrusion rates and rapid accumulation of the lava sequence. Equally however, scarcity may reflect relative susceptibility to Caledonide tectonism. The Carreg Gybi Member (Map 6), is atypical in this respect due to its relative completeness (Map 6). The Member reflects deposition during a prolonged periods of quiescence from active volcanism and the return to mudstone deposition and the background conditions of the Arenig-Llanvirn Shales {4.6.}.The Member exhibits a complex interrelationship of the products of basic and acidic volcanism, volcaniclastics deposition, high-level intrusion, and marine sedimentation (Fig. 5.3.)

The base of Carreg Gybi Member is taken, at the base of a thick (≈ 13m) pyroclastics debris flow deposits (Unit C on Map 6, Plate 5.14.) which rest on pillow lavas, basaltic hyaloclastites

Figure 5.3. Schematic sedimentary and lithological log of the Carreg Gybi Member.

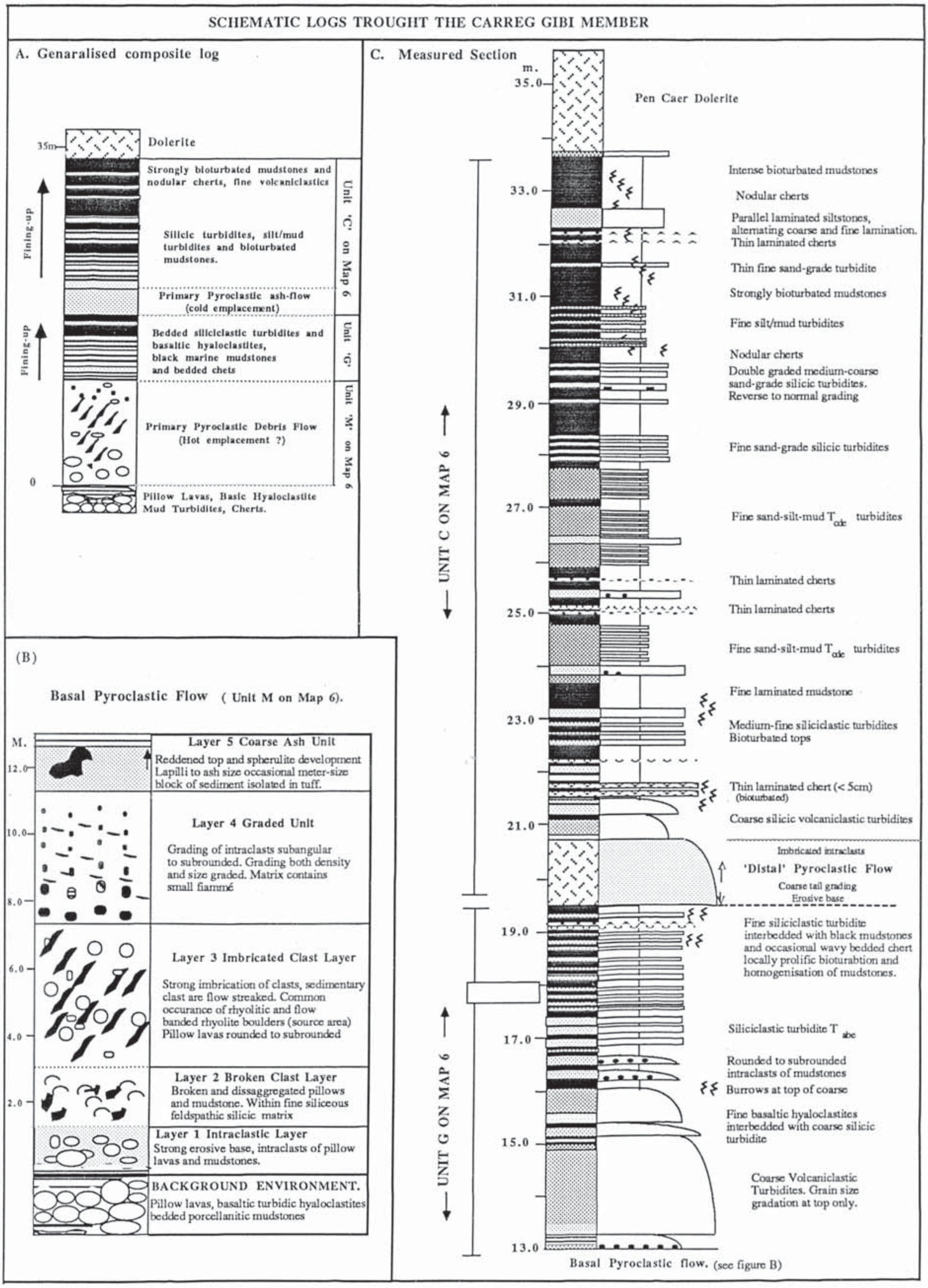
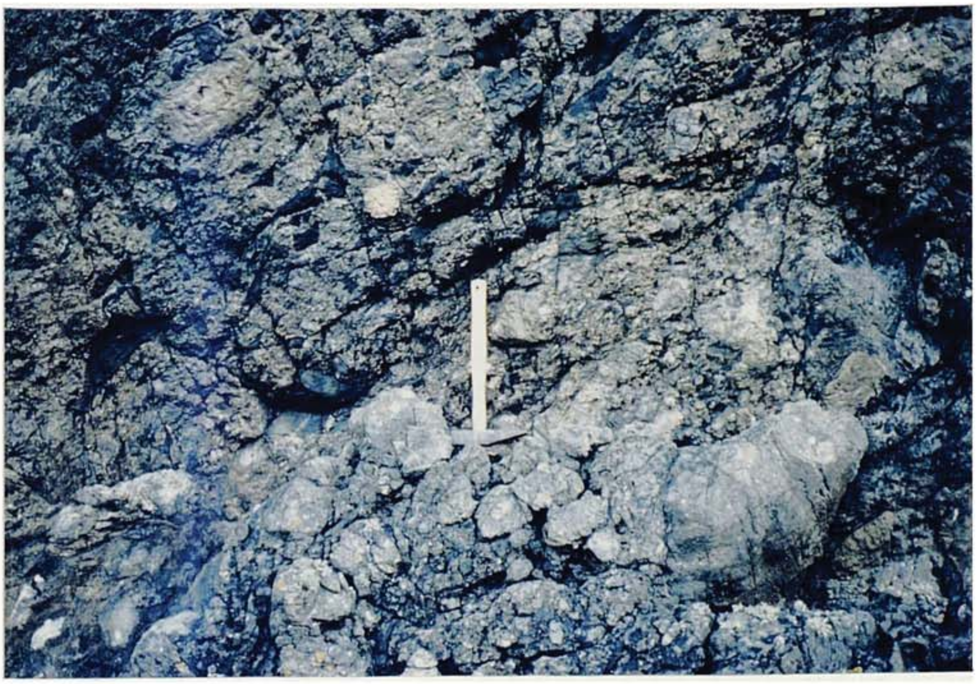
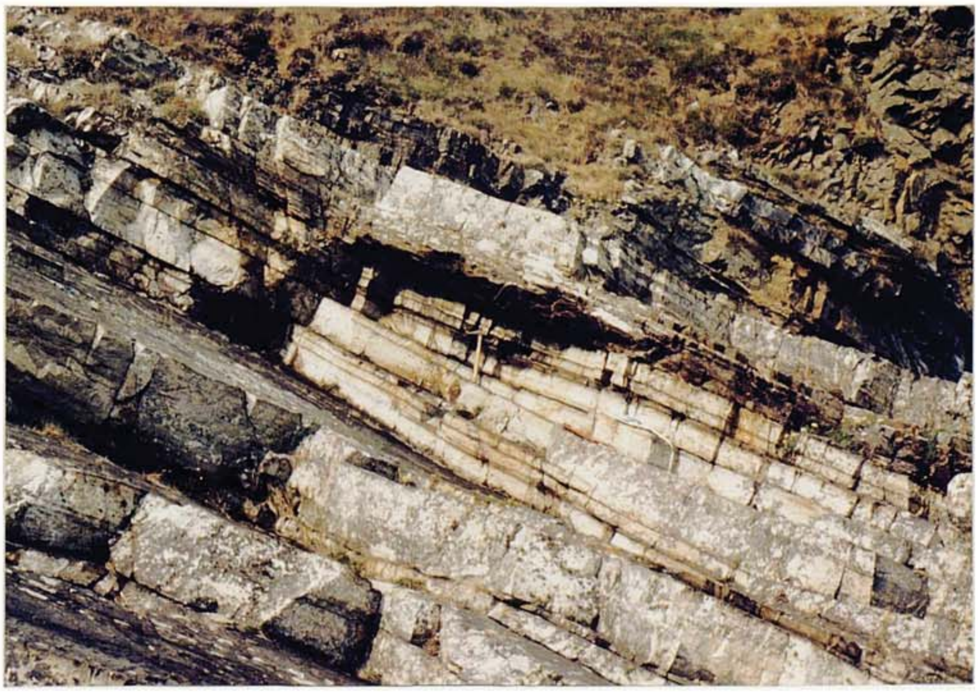

SCHEMATIC LOGS TROUGHT THE CARREG GIBI MEMBER
A. Genaralised composite log
35m
Dolerite
Strongly bioturbated mudstones and nodular cherts, fine volcaniclastics
Silcic turbidites, silt/mud turbidites and bioturbated mudstones.
Primary Pyroclastic ash-flow (cold emplacement)
Unit 'C' on Map 6
Fining-up
Bedded siliciclastic turbidites and basaltic hyaloclastites, black marine mudstones and bedded chets
Unit 'G'
Fining-up
Primary Pyroclastic Debris Flow (Hot emplacement ?)
Unit 'M' on Map 6
0
Pillow Lavas, Basic Hyaloclastite Mud Turbidites, Cherts.
(B)
Basal Pyroclastic Flow (Unit M on Map 6).
M.
12.0
10.0
8.0
6.0
4.0
2.0
Layer 5 Coarse Ash Unit
Reddened top and spherulite development Lapilli to ash size occasional meter-size block of sediment isolated in tuff.
Layer 4 Graded Unit
Grading of intraclasts subangular to subrounded. Grading both density and size graded. Matrix contains small fiammé
Layer 3 Imbricated Clast Layer
Strong imbrication of clasts, sedimentary clast are flow streaked. Common occurance of rhyolitic and flow banded rhyolite boulders (source area) Pillow lavas rounded to subrounded
Layer 2 Broken Clast Layer
Broken and dissaggregated pillows and mudstone. Within fine siliceous feldspathic silicic matrix
Layer 1 Intraclastic Layer
Strong erosive base, intraclasts of pillow lavas and mudstones.
BACKGROUND ENVIRONMENT.
Pillow lavas, basaltic turbidic hyaloclastites bedded porcellanitic mudstones
C. Measured Section
m.
35.0
33.0
31.0
29.0
27.0
25.0
23.0
21.0
19.0
17.0
15.0
13.0
Pen Caer Dolerite
UNIT C ON MAP 6
UNIT G ON MAP 6
Intense bioturbated mudstones
Nodular cherts
Parallel laminated siltstones, alternating coarse and fine lamination.
Thin laminated cherts
Thin fine sand-grade turbidite
Strongly bioturbated mudstones
Fine silt/mud turbidites
Nodular cherts
Double graded medium-coarse sand-grade silicic turbidites. Reverse to normal grading
Fine sand-grade silicic turbidites
Fine sand-silt-mud T_{cde} turbidites
Thin laminated cherts
Thin laminated cherts
Fine sand-silt-mud T_{cde} turbidites
Fine laminated mudstone
Medium-fine siliciclastic turbidites Bioturbated tops
Thin laminated chert (< 5cm) (bioturbated)
Coarse silicic volcaniclastic turbidites
Imbricated intraclasts
'Distal' Pyroclastic Flow
Coarse tail grading
Erosive base
Fine siliciclastic turbidite interbedded with black mudstones and occasional wavy bedded chert locally prolific bioturabtion and homogenisation of mudstones.
Siliciclastic turbidite T_{abc}
Rounded to subrounded intraclasts of mudstones
Burrows at top of coarse
Fine basaltic hyaloclastites interbedded with coarse silicic turbidite
Coarse Volcaniclastic Turbidites. Grain size gradation at top only.
Basal Pyroclastic flow. (see figure B)

Plate 5.13. Basaltic hyaloclastic turbidites from the Strumble Head Volcanic Formation underlying the Carreg Gybi Member The tectonic veins are comprised of prehnite.

Plate 5.14. Pyroclastic debris flow forming the base of the Carreg Gybi Member. The base of the flow is just below the field of view. The hammer in center view is resting on basaltic pillow lavas which are incorporated within the flow. Above this, note the imbrication and streaking of intraclastic debris (see Fig 5.3.)

Plate 5.15. Silicic volcaniclastic turbidites from the Carreg Gybi Member.





flow deposits (Unit C on Map 6, Plate 5.14.) which rest on pillow lavas, basaltic hyaloclastites and marine shales (see Plate 3.3.). The *pyroclastic flow* is characterised by the abundant intraclastic debris incorporated from the underlying beds. The flow possess, discrete layering, sorting, and a fabric, in character with documented subaqueous pyroclastic deposit (Fiske & Matsuda 1964, Yamada 1984). It is thought that this deposit may have been emplaced whilst still hot. The erosive properties of the flow, whilst unduly evident (Plate 5.14.), are seen to be variable across its base. Impersitant selvedge of turbidic mudstones are seen on the headland of Carreg Gybi to the east, whilst to the west at Pen Caer (approx 100m) the flow rests on 2m of mudstones and basaltic volcaniclastics.

Whilst the base of the Member is represent by deposition from a single event pyroclastic flow, the background sedimentation was essentially mudstone (Fig. 5.3.). The accumulation of mudstone in such thicknesses, probably indicates that extrusive basaltic activity ceased for a relatively prolonged period. However, a shift in the site of active extrusion to elsewhere in the Fishguard Volcanic Complex may be appropriate, given that basaltic hyaloclastic tuffs are a small although significant component of the sedimentary sequence.

The nature of the fine grained sediments can be seen to vary from homogenous black shales, to faintly parallel laminated mudstone-siltstone, bedded and nodular cherts, and cross-laminated coarse siltstones and fine sandstones. The sediments represent predominantly pelagic sedimentation, from suspension and low-density low-velocity turbidity currents. Fine silt grade parallel lamination predominates throughout the succession, although locally the mudstones are structureless, possibly due to homogenisation by burrowing. Bioturbation of the pelitic sediments is highlighted by preferential weathering on exposed surfaces, mixing of lithologies, and disruption of lamination, whilst preferential prehnitisation of burrows is locally evident. Thin bedded (<5cm), non-erosive and graded fine sandstone-siltstone-mudstone beds occur frequently throughout the Member. They generally posses evidence of small scale asymmetrical ripples, and are thought that they represent T_{cde} turbidites, appearing similar to low-velocity low-density current-rippled siltstone turbidites described by Pickering (1984) from the Båsnaering Formation, Norway.

Planar bedded chert horizons occur sporadically within the argillic beds (Fig. 5.3.), readily identifiable by their conspicuous semi-translucent colour. Such horizons vary in thickness from 1-4cm, traceable throughout outcrop and showing little lateral variation, although locally, bioturbated results in a wavy bed forms of intermixed silica and mudstone. Bioturbation of the cherts indicates that they were probably amorphous silica, and represented a soft silicous sediments which accumulated on the sea-bed. Nodular cherts occur at discrete horizons, generally at the top of the Member (Fig. 5.3.). Nodules vary in size from several centimetres to tens of centimetres, composed of dull-grey chalcedonic silica. The shape of nodule is invariably spheroidal, although occasionally a flattened disc appearance is observed. The occurrence of both shapes within individual beds, suggests that shape is a primary feature and not tectonic modification, with their origin as diagenetic concretions which formed below the sediment-water interface.

Interbedded with the pelitic sediments, thick silicic volcaniclastics turbidites are well developed (Plate 5.15.). The nature of bedding within the silicic volcaniclastics is essentially one of massive bedded units varying in thickness from 20-70cm. The silicic turbidites are essentially moderately sorted quartz-feldspar sandstones. Lower contacts are commonly erosive bases and mudstone 'rip-ups' are frequent. They locally show evidence for a variety of sedimentary structures, including symmetric grading (reverse to normal). Bioturbation is common to the top of individual beds. Interbedded within the silicic volcaniclastics and pelitic sediments, a further single event pyroclastic flow is developed (Fig. 5.3). The deposit is poorly sorted with a tuffaceous matrix differing from the silicic volcaniclastics turbidites due to its textural and compositional immaturity.

Summary of basaltic volcanism

Basaltic volcanism was generally effusive in nature, pillowed lavas and tube networks volumetrically dominate. However, sheet flows, pillow breccia, high level contemporaneous intrusives, bedded basaltic hyaloclastites, and basic volcaniclastics developed locally. The deposits represent a complex interplay of extrusive, hydroclastic, autoclastic and gravity driven processes, within the subaqueous environment. Volcanism was probably located along small fissures, only in rare instances did topographic highs developed. The presence, of interbedded pelitic and porcellanitic mudstones, and various volcaniclastics, deposited in the main by turbidity currents, would support the contention of Kokelaar *et al.* (1984a) that basaltic volcanics are unlikely to have formed any marked topographic feature on the sea-floor, with the basaltic lavas developed within a 'graben-like' depression, within which subsidence kept pace with the extrusion of lava.

5.5. SUMMARY

Prior to the development of the Fishguard Volcanic Complex, there is evidence to indicate that volcanism and intrusive activity had occurred within the Fishguard Porth-gain region during late Arenig times. This activity is represented the Yns-Castell Ash (*c.f.* Cox 1916) and Pwll Deri Ashes (Map 1.). The Yns Castell Castle Ash are a sequence of subaqueous pyroclastic debris flows. The Pwll Deri Tuff is newly recognised and contains a varied range of lithologies, which include welded ash-flow tuffs, suggesting the presence of a small volcanic centre near the Fishguard region, prior to the development of the Fishguard Volcanic Complex.

The presence of igneous activity during late Arenig times is supported by the identification of intrusives, the age of which is supported by the identification of Arenig intrusive activity.

Upper Llanvirn igneous activity is recorded by the Fishguard Volcanic Group. Stratigraphically the Group is difficult to interpret in terms of a tripartite volcanic assemblage as documented by previous workers. It is however, the simplest solution with the limited exposure and the identification of repetition within well defined successions (Map 5-6). One modification to the Group is the introduction of the Carreg Gybi Member. The environmental interpretation of the Volcanic Group, of small rhyolite domes, and a suite of pillow lavas erupted within a graben like structure (Bevins 1982, Kokelaar *et al.* 1984.a.), appears to be an appropriate setting.

CHAPTER SIX
METAMORPHIC GEOLOGY

6.1. INTRODUCTION

It is now widely accepted that the Welsh Caledonides have been variably affected by low grades of metamorphism (e.g. Bevins & Rowbotham 1983, Roberts & Merriman 1984). The secondary mineralogy of basic lithologies throughout this study area allows a quantitative evaluation of the metamorphic conditions which prevailed in the Fishguard district during this regional event (or events). In an attempt to establish the conditions mineral assemblage development is discussed, and the chemistry, form and paragenesis of individual phases described. A brief discussion regarding the secondary mineralogy of fault-rock products, silicic, and sedimentary rock types is given. Detailed literature reviews regarding the historical development of metamorphic studies within the Welsh Basin are given by Bevins (1979) and Cronshaw (1985), and are not discussed here. A review of the facies concept in low-grade metamorphic terrains is more than adequately covered by Coombs (1960) and Turner (1981).

To establish the metamorphic conditions of the Fishguard district in terms of P-T, it is convenient, throughout much of the following chapter, to discuss metamorphism as having been under the singular influence of normal burial processes. It is worth noting at this point, that metamorphism and phase development reflects a complex interaction of several factors, including, variation in fluid chemistry, precursor mineral chemistry, and Caledonide tectonism; whilst it is likely that thermal gradients fluctuated and burial history changed through time.

6.2. MINERAL PHASES AND MINERAL ASSEMBLAGE DEVELOPMENT

An outline of the primary igneous mineralogy and textural development of the various basic lithologies is given in Chapter 5. The phases clinopyroxene, plagioclase feldspar and ore (Fe-Ti oxides) are discussed here in the context of their 'relict' relationship regarding their control on secondary phase development during subsequent metamorphism. Associated with these relict minerals, basic lithologies can be discussed in terms of their secondary mineralogy regarding fifteen phases (+ quartz). Eleven of these phases (actinolite, albite, calcite, chlorite, epidote-clinozoisite, prehnite, pumpellyite, ferri– and ferrostilpnomelane, sphene, white mica, and minor K-feldspar) are thought to have developed largely as a consequence of low-grade regional metamorphism. The remaining four phases are amphiboles (actinolitic hornblende, winchite, hornblende, kaersutite) which appear to have developed prior to regional low-grade metamorphism and are referred to here as *'pre-regional'* phases. Causative reasons for their development is briefly discussed {6.3.} and chemical properties outlined {6.4.4.}; their peripheral interest precludes detailed discussion.

Plagioclase is invariably albitised, although compositional variation is locally apparent (An 0-28 mol%). Clinopyroxene is relatively unaltered in both extrusive and intrusive lithologies, its inert nature being typical of low-grade metabasites. Chlorite is pervasive its occurrence is attributable in part to a sideromelane/palagonite precursor in extrusives and areas of interstitial mesostasis in intrusives. Sphene is common, associated with interstitial chlorite and replacing the ilmenite fraction of ore, frequently leaving magnetite unaltered. Stilpnomelane has been observed in both iron oxidation states, more typically as ferristilpnomelane. White mica is generally observed replacing plagioclase feldspar, whilst K-feldspar is conspicuous by its scarcity. Calcite is generally late in the paragenetic sequence and, where abundant, Ca-Al silicates are commonly absent.

Whilst the above are important with regards the general character of the observed mineral assemblages (and by modal volume make up at least 75% of all polymineralic assemblages - excluding metadomains), of specific interest is the development within a given assemblage, of the 'index' phases actinolite, prehnite, epidote, and pumpellyite. The problem when assessing mineral assemblage development is accentuated by the apparent lack of agreement of which scale equilibrium should be sought, i.e.; phases in contact (Kawachi 1975), less than four millimeters (Houghton 1982), or phases within an area of a standard thin-section (Coombs *et al.* 1976). Recently, Evarts & Schiffman (1983) in a detailed study of the 'Del Puerto Ophiolite' central California, whilst identifying prevalent microdomain disequilibria, suggest that predictable mineralogical changes in the context of the ophiolite's stratigraphy indicate that in general equilibrium had been attained. The proposals of Kawachi (1975) and Houghton (1982) whilst meritorious, are relatively unrealistic in coarser grained rock types; while the decimetre scale of Evarts & Schiffman (1983) is not applicable due to the stratigraphic complexity. The proposal of Coombs *et al.* (1976), has therefore been favoured in this study, although the reservations of using such a scale is acknowledged (see Vernon 1976).

6.2.1. Metabasite index assemblages (*inc.* Metadomains)

The Fishguard metabasites contain four assemblage types that are diagnostic of a specific low-grade facies, and which can be regarded as 'index' assemblages in the context of low-grade metamorphism.

The index assemblages are; (+ chlorite + albite + sphene + quartz ± others)

1) pumpellyite + epidote 2) pumpellyite + prehnite 3) pumpellyite + epidote + prehnite	**prehnite-pumpellyite facies assemblages**
4) actinolite + epidote	**greenschist facies assemblage**

A fifth assemblage termed here a *'transitional assemblage'*, is also widely developed in the Fishguard district. Transitional assemblages may contain all four index phases in variable proportions, although not in true equilibrium; however, the sub-assemblage actinolite+epidote is invariably in textural equilibrium,

5) epidote + actinolite ± prehnite ± pumpellyite **'transitional' assemblage**

Metadomains (*c.f.* Smith 1968b) characterised by pervasive alteration, limited numbers of secondary phases and where the primary mineralogy/texture is only rarely preserved, are seen locally within the larger gabbroic intrusions. They are predominantly epidote rich (i.e. epidote metadomain) and whilst their origin in some cases remains equivocal, such a style of alteration is indicative of a high fluid - rock ratio (Jolley & Smith 1972)

6) ± clinopyroxene* + epidote + chlorite ± actinolite **epidote metadomain assemblage**

Areas where index assemblages (*inc.* metadomains) were found to be common, are plotted on an outline map which depicts the spatial distribution of basic lithologies (Fig. 6.1.). Due in part to inadequate exposure and the complex volcanostratigraphy isograd delineation has not been attempted. A broad spatial distribution nevertheless appears to exist between assemblage types, whilst there development indicates that the area can not be regarded as a monofacies prehnite-pumpellyite terrain as has previously been the case (Bevins 1978, Bevins & Rowbotham 1983).

Prehnite-pumpellyite facies assemblages

Prehnite-pumpellyite assemblages are restricted in general to the pillow lavas and high level dolerites on the northern part of the Pen Caer peninsular (Fig. 6.1.). The fine grained nature of many lithologies in this area, inhibits detailed petrographic observation and hinders microprobe analysis. However, the most frequently observed index assemblages are; (+quartz ± calcite ± white mica, *denotes primary relic)

± clinopyroxene* + albite + chlorite + sphene + epidote + pumpellyite + prehnite
± clinopyroxene* + albite + chlorite + sphene + pumpellyite + prehnite

,whilst non-index assemblages include; (+ quartz ± stilpnomelane)

albite + chlorite + sphene + prehnite ± white-mica ± epidote ± calcite
albite + chlorite + calcite ± sphene ± white-mica.

The paragenesis of diagnostic assemblage development is difficult to assess. It is likely that the observed prehnite-pumpellyite assemblages (excluding metadomain assemblages {6.2.2.})

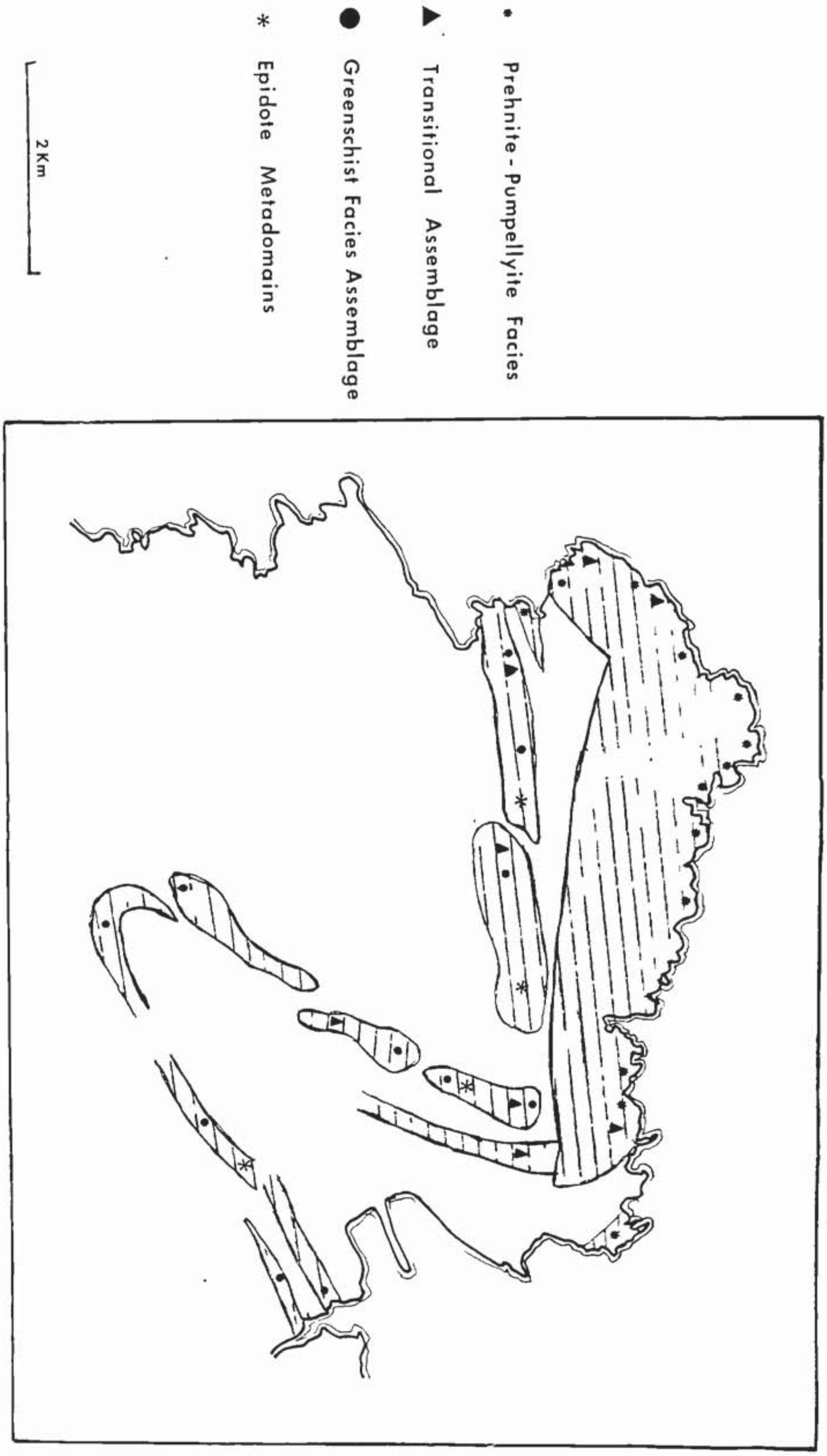


Figure 6.1. Outline map of basic lithologies within the Fishguard Volcanic Complex west of Fishguard, showing the distribution of the various mineral assemblages, including diagnostic assemblages and epidote-metadomains.

owe in part their development to a precursor zeolitic mineralogy; the absence of which indicates that the conditions of the zeolite facies was surpassed. It is plausible that specific styles of alteration, such as the widespread prehnitisation of hyaloclastites on the Pen Caer Peninsular, owe assemblage development to progressive burial, in as much as such lithologies are likely to be favoured by initial zeolitic growth by virtue of their high comparative permeability. The frequent occurrence of quartz in prehnitised hyaloclastites may owe development to dehydration reactions such as;

$$CaAl_2Si_4O_{12} . 4H_2O + 2(OH)^- = Ca_2Al_2Si_3O_{10} (OH)_2 + SiO_2 + H_2O$$

(i.e. laumontite = prehnite + quartz + H_2O, *c.f.* Houghton 1982).

Whilst prograde burial may be a probable progression to the entire volcanic pile, within basaltic cataclastites and various fault-rock products {3.4.} throughout the upper parts of the Strumble Head Volcanic Formation, prehnite and pumpellyite can be observed to have crystallised directly from the fluid phase. This is analogous in part to hydrothermal ocean floor type metamorphism; although of course the fluid here is tectonically introduced at the time of Caledonide deformation.

Whilst prehnite-pumpellyite index assemblages are widely developed throughout the upper parts of the Strumble Head Volcanic Formation, equally prevalent are non-diagnostic Ca-Al silicate assemblages; good examples of which are developed around the headland of Carreg-wasted Point [S.M. 9260 4060] and subjacent areas, where the dominant assemblage is one of,

calcite + albite + chlorite ± white-mica ± sphene

with Ca-Al index silicates absent. The suppression of Ca-Al silicates by calcite is well documented, attributed to high μCO_2 in the fluid phase (e.g. Glassey 1974). However, the majority of calcite is demonstrably late in the paragenetic sequence occurring in cross-cutting fractures and veins, and is not considered to represent high μCO_2 during prograde conditions. The widespread development of prehnite attests to this, as a very low μCO_2 is required for its crystallisation (Thompson 1971).

'Transitional' type assemblages

The term transitional assemblage is used here where the equilibrium sub-assemblage conforms to that of either a greenschist (actinolite + epidote + chlorite) or a non-diagnostic epidote assemblage (epidote + chlorite); although the total assemblage may contain prehnite or pumpellyite, or both, as variegated phases in obvious disequilibrium. Transitional assemblages, whilst difficult to interpret in terms of model equilibrium are of interest, in that they possibly represent assemblages which developed close to the subgreenschist - greenschist facies transition; the P-T conditions of which are well defined by experimental phase relations.

Transitional assemblages are mineralogically variable. In an attempt to show such diversity three representative examples are discussed below; examples are taken from differing localities and lithologies, namely; *a)* pillow breccias - Pwl Arian, *b)* basic tuffs - Carn Hendi, *c)* gabbros - Llwanda.

a) In the pillow breccias at Pwl Arian [S.M. 8855 4035], the dominant assemblage is comprised of the following; (+ quartz)

chlorite + albite + sphene + epidote + actinolite + prehnite ± pumpellyite ± K-feldspar

Actinolite is a minor component, although present as fine bi-pyramidal rhombs in close association with prismatic epidote within a chloritic groundmass (Plate 6.2.); this is the textural equilibria sub-assemblage. Conversely, prehnite and pumpellyite occur as highly corroded spongy aggregates with no discernible crystal outlines, strongly indicating breakdown and disequilibria. Whilst reactions have obviously not gone to completion, it is thought that the observed sub-assemblage may represent development by one of the following;

pumpellyite + chlorite + quartz = actinolite + epidote + H_2O
prehnite + chlorite + quartz = actinolite + epidote + H_2O

Such reactions are important as they represent two of the major actinolite producing reactions in low-grade metabasites, having been used as key reactions to experimentally define the sub-greenschist - greenschist facies transition at low pressures (Nitsch 1971, Liou *et al.* 1985).

b) Within the pillow lavas and basic tuffs at Carn Hendi [S.M. 9365 3890] epidote, prehnite and pumpellyite assemblages predominate. They differ however, from prehnite-pumpellyite equilibria assemblages on the Pen Caer peninsular in as much as prehnite and pumpellyite show widespread evidence for textural breakdown. The equilibria sub-assemblage is; (+ quartz)

chlorite + epidote + albite + sphene ± white-mica ± calcite

whilst the total assemblage may comprise,

chlorite + albite + sphene + epidote ± prehnite ± pumpellyite ± white-mica ± calcite

Idiomorphic epidote may locally be observed to radiate from corroded aggregates of pumpellyite, suggesting it is forming at its expense; the reaction for which entails, minimal cation exchange, dehydration, and oxidation within a fluid phase (Jolley & Smith 1972), e.g.;

$$(\text{Pumpellyite})\ Ca_4(Mg_{0.5}\,Fe^{2+}{}_{0.5}\,Fe^{3+}{}_{1.0})\,Al_4Si_6O_{21}\,(OH)_7 + 0.5\,Fe^{2+} + 0.25\,O_2 \Longleftrightarrow$$

$$2(Ca\,(Al, Fe^{3+})\,Si_3O_{12}\,(OH)) + 0.5\,Mg^{2+} + 2.5\,H_2O\ (\text{Epidote 33 mol\% Ps})$$

Such 'epidote-dominated' assemblages are difficult to interpret, although it is thought realistically to regard them as *'transitional'* between the prehnite-pumpellyite and greenschist facies, in as much as prehnite and pumpellyite are relict with regard to the equilibrium sub-assemblage. Whilst it is obvious that reactions have not gone to completion and actinolite is absent, it is likely that conditions were approaching, or had reached, the sub-greenschist - greenschist facies boundary. It is of interest to note that epidote bearing actinolite-free assemblages have been regarded as greenschist facies assemblages characteristic of low-pressures under high fO_2 (Gass & Smewing 1973). Conditions which are not too dissimilar from those thought to have existed in the Fishguard district {6.6}. (*n.b.* pumpellyite consuming reactions are not thought to reflect a zeolite/prehnite-pumpellyite facies transition).

c) In the aforementioned examples of transitional assemblages, pumpellyite and particularly prehnite, whilst in disequilibria, are present in significant quantity. In the gabbros of Llwanda and Carn Gelli [S.M. 9330 3960] these phases are scarce, although nevertheless present in small amounts. Both prehnite and to a lesser extent pumpellyite exist as minor subordinate relic remnants in what may otherwise be regarded as greenschist facies assemblages. Prehnite is observed as fine variegated flakes within feldspar and more rarely within interstitial chlorite, where it may be seen to be replaced by prismatic epidote and acicular actinolite; which may relate to a prehnite consuming reaction outlined above (i.e. prehnite + chlorite + quartz = actinolite + epidote + H_2O). Typical assemblages may be complex due to the presence of 'pre-regional amphiboles', although may be comprised of the following; (+ quartz, # denote 'pre-regional' amphibole)

clinopyroxene* ± kaersutite# ± blue-green actinolitic hornblende# + albite + chlorite + epidote + sphene + actinolite + prehnite ± pumpellyite + calcite ± white mica

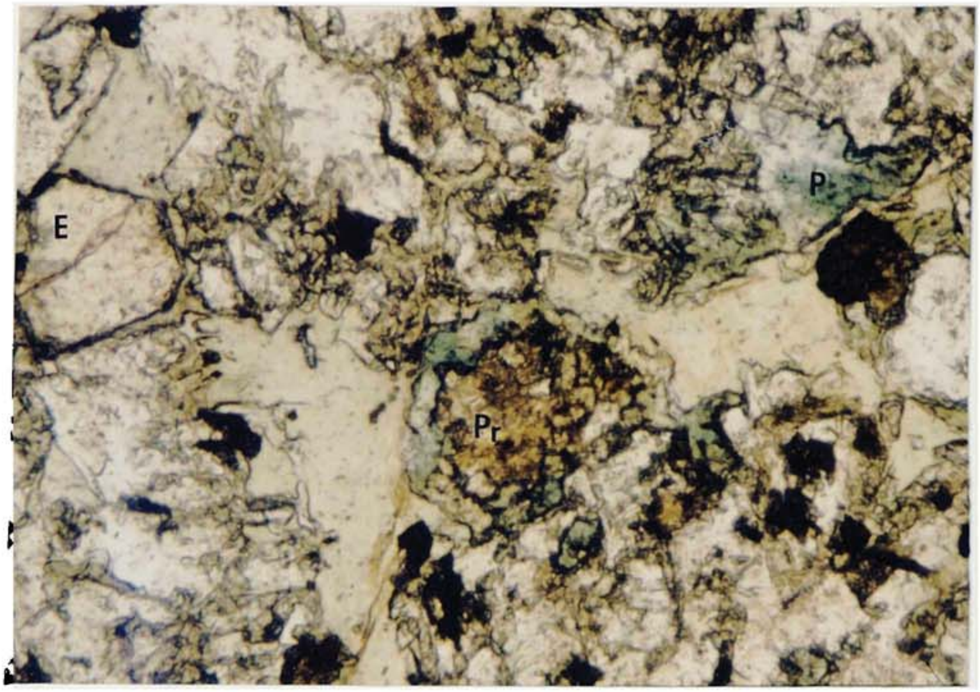
Pumpellyite is rare in the gabbros of the Llwanda and Garn Fawr districts, this is at variance to the findings of Bevins (1978) who suggest that it is common, and widely developed.

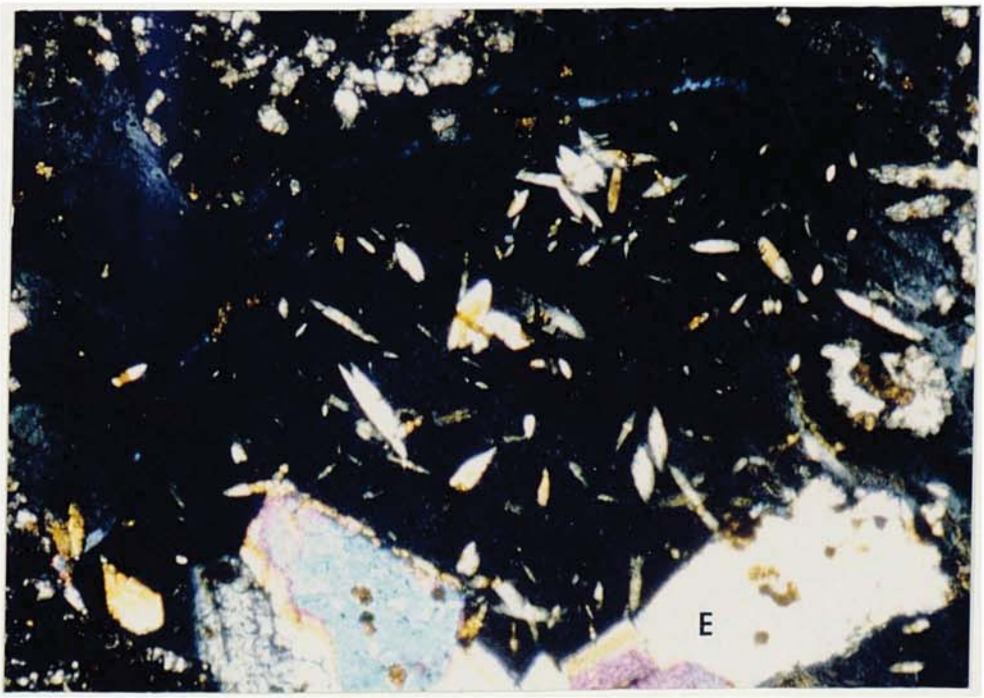
Greenschist facies assemblages

Greenschist facies assemblages dominate the metabasites from Lower Town Fishguard through Manorowen, Carn Gelli, Llwanda, Garn Fawr, and towards Porth Maenmelyn (Fig. 6.1.). Transitional type assemblages are present in the latter three districts, although typical greenschist assemblages are also widely evident. It is reasonable to suggest that conditions of greenschist facies had been attained in these districts, the presence of typical lower grade relic

Plate 6.1. *Prehnite - Pumpellyite facies assemblage.* Photomicrograph of a typical prehnite-pumpellyite facies assembalge developed in a dolerite from the Carreg Gybi headland (Map 6.). The assemablage comprises, subhedral aggregates of pumpellyite (P: in part replacing feldspar), sphene (high-relief, dark), tabular epidote (E), and dusty aggregates of prehnite (Pr: surrounded by pumpellyite), set within a chlorite groundmass (low relief) which may reflect areas of interstial mesostasis. Field of view 0.89mm.

Plate 6.2. *Transitional assemblage.* Photomicrograph of a transitional assemblage from pillow breccias at Pwl Arian. The dark background is chlorite, within which fine by-pyramidal rhombs of actinolite are developed. Tabular equant epidote (E) is seen at base of plate, traces of sphene are present throughout. The spongy corroded aggregates in top centre view is prehnite intermixed with traces of K-feldspar. Pumpellyite is not seen in the field of view although is present in the thin section as highly corroded aggregates. Field of veiw 0.35mm.


E
P
Pr


E

phases possibly indicating slow reaction rates. There is no evidence for the retrogression of greenschist assemblages. In the aforementioned districts, assemblages typically comprise the following; (+ quartz + chlorite ± white-mica)

± clinopyroxene* + albite + sphene + actinolite + epidote ± stilpnomelane ± calcite

Such an assemblage (Plate 6.3.) can be regarded as diagnostic of the greenschist facies. Bevins & Rowbotham (1983), indicate that pumpellyite bearing assemblages dominate the aforementioned areas around Fishguard, although no evidence has been found from the specimens examined in this study to support this.

Textural evidence within transitional assemblages indicates that prehnite and pumpellyite are in disequilibria, and breaking down to form actinolite and epidote. This may suggest that greenschist assemblages owe in part their development under progressive conditions to reactions such as;

$$32\ Ca_2Al_2Si_3O_{10}(OH)_2 + 5\ Mg_4Al_4Si_2O_{10}(OH)_8 + 10\ SiO_2 \longleftrightarrow$$
$$28\ Ca_2Al_3Si_3O_{12}(OH) + 4\ Ca_2Mg_5Si_8O_{22}(OH)_2 + 34\ H_2O$$

(prehnite + chlorite + quartz = epidote + actinolite + H_2O)

$$9\ Ca_4Al_5MgSi_6O_{21}(OH)_7 + Mg_6\ Si_4O_{10}(OH)_8 + 11\ SiO_2 \longleftrightarrow$$
$$3\ Ca_2Mg_5Si_8O_{22}(OH)_2 + 15\ Ca_2Al_3Si_3O_{12}(OH) + 25\ H_2O$$

(pumpellyite + antig. mol. of chlorite + quartz = tremolite + clinozoisite + H_2O)

Whilst evidence for actinolite + epidote + stilpnomelane assemblages in close proximity to calcite + quartz veins may suggest that reactions such as;

$$3\ Mg_3Al_2Si_3O_{10}(OH)_8 + 10\ CaCO_3 + 21\ SiO_2 =$$
$$3\ Ca_2Mg_5Si_8O_{22}(OH)_2 + 2\ Ca_2Al_3Si_3O_{12}(OH) + 8\ H_2O + 10\ CO_2$$

(chlorite + calcite + quartz = actinolite + epidote + H_2O + CO_2)

were locally operative in producing greenschist assemblages. The absence of hornblende, lack of evidence for a peristerite gap (Maruyama *et. al.* 1983), or an increase in the anorthite content of plagioclase within greenschist facies assemblages, strongly indicates that conditions of the amphibolite facies were not attained (i.e. T > 380°C, Fig. 6.11.).

Metadomain assemblages

Metadomains are developed locally throughout the Fishguard Volcanic Complex, particularly within the gabbroic intrusions which underlie the Fishguard Volcanic Group forming a discontinuous intrusive belt from Garn Fawr to Manorowen (Fig. 6.1.). The gross nature of metadomains are variable, although two major types can be distinguished; *epidote-metadomains* and *tectono-metadomains*; the former having obvious mineralogical connotations, the latter generic.

Epidote Metadomains: Throughout many of the large gabbroic intrusions epidote-metadomains can be observed. The style, shape, and size of which may vary from 'patch-like' to 'veined' networks; replacing small areas in isolated occurrence to areas of metre size scale readily recognisable at outcrop due to a lime-green waxy colouration and common preservation of relic texture. In thin-section epidote-metadomains are dominated by piscasitic epidote, chlorite is scarce, whilst quartz rich vugs are numerous. Epidote habit is variable from large euhedral crystals to amorphous interlocking plates. Feldspar and ore is generally replaced, although clinopyroxene may be preserved as heavily corroded relics, commonly pseudomorphed by epidote platelets and less frequently by acicular actinolite. This common assemblage may indicate conditions of alteration are of the greenschist facies, however, the style of alteration is not comparable with a 'regional' greenschist facies assemblage development, it being likely that separate events are responsible.

By their nature of alteration metadomains are commonly developed in areas of high comparative permeability, such as vesicular margins of lava flows and in the vicinity of faults. Whilst this may readily account for 'vein-type' epidote-metadomains, there appears no obvious pathway to increase permeability in the development of 'patch-type' epidote-metadomains, consequently the origin of which remains uncertain. They are however, commonly observed in coarse grained and pegmatitic areas which may suggest that alteration is by late stage deuteric fluids. Equally plausible is a method of epidotisation comparable to that observed in ocean-floor gabbros, where alteration is associated with hydrothermal circulatory systems (e.g. Cann 1979). A similar process has been speculatively accredited to pumpellyite metadomains within the Builth Volcanic Group, where an inferred igneous complex or magma chamber at depth drives a localised hydrothermal system (Bevins 1985). In the Llwanda and Carn Gelli districts there is field evidence for post-gabbroic felsic intrusion which presumably had a source at depth {5.1.}; a similar process may therefore realistically be envisaged.

Tectono-metadomains: Tectono-metadomains are areas of alteration derived by increased permeability as a direct consequence of tectonism. Pervasive alteration is generally restricted to cataclastic debris within the fault zone, although may be observed to permeate outwards into the undeformed host rock. It appears that the size of the faults has a comparatively limited control on alteration, it being the fracture density which dictates the scale of alteration. For example thrusts (which are by far the major structures in controlling regional geology), have in general, a low fracture density outside of the main plain of dislocation, with alteration restricted to the thrust plain and peripheral margins. Many cross-cutting Caledonide trending normal faults however, whilst of extrinsic value, exhibit essentially brittle behavior with high fracture density throughout the surrounding wall rock, conversely exhibiting widespread alteration. The mineralogy of tectono-metadomains is discussed below, whilst their deformatory mechanisms are outlined in Chapter 3.

6.2.2. Alteration and amphibole development ('pre - regional alteration').

It is highly probably by analogy with modern oceanic basalts (e.g. Baragar *et al.* 1977, Mevel 1981), that the basic lithologies throughout the area suffered chemical and physical alteration prior to low-grade regional metamorphism. Halmyrolosis, hydrothermal alteration, and palagonitisation are likely to have been operative, although evidence for such processes is generally lacking. Bevins & Roach (1979) have suggested that potassium rich globules observed in the margins of rhyodacite pillow lavas may reflect low-temperature sea-water alteration. The best preserved evidence for early alteration, resulting possibly by the hydrothermal venting of silica, is the intermittent association of chert and pillow lavas on the Pen Caer peninsular {5.4.}. It is realistic to assume however, that features such as the chloritisation of hyaloclastites is almost certainly the end product of early divitrification and palagonitisation.

Within the gabbroic intrusions of the district there is considerable petrographic evidence for the local development and existence of secondary amphiboles, prior to low-grade regional metamorphism. It is apparent that such amphiboles are the consequence of two separate processes; one akin to post-magmatic low-pressure amphibolitisation responsible for the development of a hornblende + quartz assemblage, the other of restricted occurrence associated with felsic micro-dykes producing an amphibole assemblage characterised by sodic-calcic amphiboles (winchite - richterite) and blue-green actnolitic-hornblende.

Evidence for low-pressure post-magmatic amphibolitisation is based on the common occurrence of secondary hornblende which invariably shows evidence for replacement by actinolite. The relationship of coexisting hornblende-actinolite pairs {6.4.2.} is not thought to represent a miscibility gap (*c.f.* Klien 1969); rather, textural relationships indicate replacement of a relic secondary hornblende by actinolite that developed under the subsequent conditions of regional metamorphism.

Texturally there can be little doubt as to the secondary nature of hornblende {6.4.}. Its common mutual interrelationship with quartz, frequently containing abundant apatite rhombs, suggests crystallisation as a product of post-magmatic hydrothermal or deuteric fluids. The unusually high ferric iron content of analysed hornblende (as inferred by Fe^{2+}: Fe^{3+} normalizing calculation), suggests oxidising conditions of crystallisation, as high fO_2 will fix Fe^{3+} in oxides or silicates (Seyfried & Bischoff 1977, Evarts & Schiffman 1983). Whilst requiring further study, hornblende paragenesis may have arisen out of post-magmatic hydrothermal circulation, possibly of sea-water which would provide a high oxygen flux (Spooner & Fyfe 1973), along fractures and permeable strata analogous to hydrothermal amphibolitisation of high level oceanic gabbros as observed from ophiolite complexes (Sievell & Waterhouse 1984). Such a process should not be singularly dismissed, as hornblende chemistry suggests low-pressure crystallisation {6.4.}, whilst the volcanostratigraphy of the Fishguard Volcanic Complex is not too dissimilar from the uppermost parts of ophiolite complexes; it seeming likely that some local convection of fluids would have been associated with the larger gabbroic intrusives.

Whilst hornblende is common to many of the larger gabbroic bodies, a more localised 'pre-regional' phase of amphibole development is suggested by the presence of a secondary assemblage developed in felsic micro-dykes which cross-cut the Llwanda and Carn Gelli gabbros. The assemblage is characterised by sodic-calcic amphiboles of the winchite-richterite series and blue-green actinolitic hornblende (and Fe*-rich equivalents). The restricted occurrence of this assemblage to felsic micro-dykes and peripheral edges of the host gabbro, strongly indicates that it results from dyke emplacement and possible Na-metasomatism.

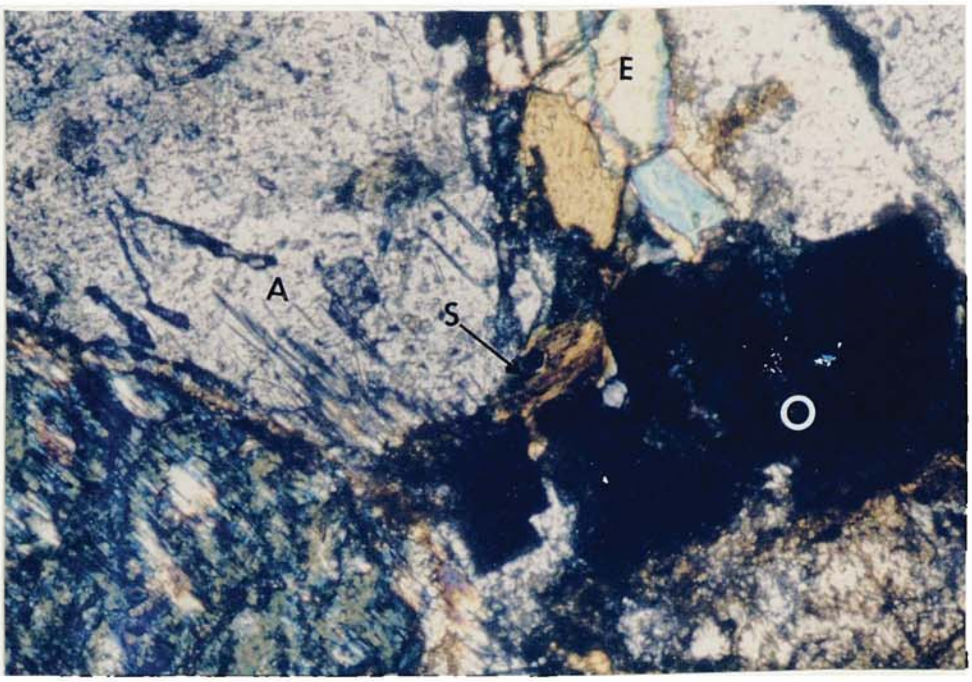
6.3. MINERALOGY OF SILICIC VOLCANICS AND FAULT-ROCKS

The development of secondary minerals is not restricted to basic lithologies of the Fishguard Volcanic Complex; chlorite, sphene, albite, stilpnomelane, Ca-Al silicates and white-mica are observed in variable proportions and assemblages in all silicic volcanics, intra-volcanic complex sediment and importantly fault-rocks associated with Caledonide deformation. Whilst such development is difficult to interpret in terms of metamorphic grade, several conclusions can be drawn from their occurrence.

Silicic volcanics and sediments: Chlorite, albite, sphene, prehnite and white-mica are variably developed in silcic and intermediate rock types on the Pen Caer peninsular. Fe-rich silicates are rare, although pumpellyite, stilpnomelane and epidote may be observed locally in tonalitic intrusions, although no index assemblages are evident. Similar non-index assemblages are widely developed within the intra-volcanic sediments, as exemplified by the heterogeneous sediments of the Carreg Gybi Member where basic volcanoclastics are pervasively altered to chlorite, sphene, minor epidote and prehnite, whilst contemporenous silicic and pelitic

Plate 6.3. *Greenschist facies assemblage.* Photomicrograph of a typical greenschist facies assemblage developed in the gabbros at Carn Gelli. The assemblage comprises fine acicular actinolite (A), epidote rhombs (E). ferro-stilpnomelane (S which is arrowed), and dusty aggregates of sphene replacing ore (O). The actinolite shows partial uralic replacement of clinopyroxene (bottam left hand corner). Field of view 1.7.mm.

Plate 6.4. *Metabasic cataclastite* Photomicrograph of a metabasic cataclastite from the Carreg Gybi dolerite. Note extensive grain size reduction of clinopyroxenes porphyroclasts (light coloured aggregates), althought they remain relatively fresh. The groundmass is composed entirely of fine grained acicular pumpellyite, whilst the coarses aggregates to the left of filed of view is prehnite. The black area top of plate is a hole in the slide. Field of view 3.8mm.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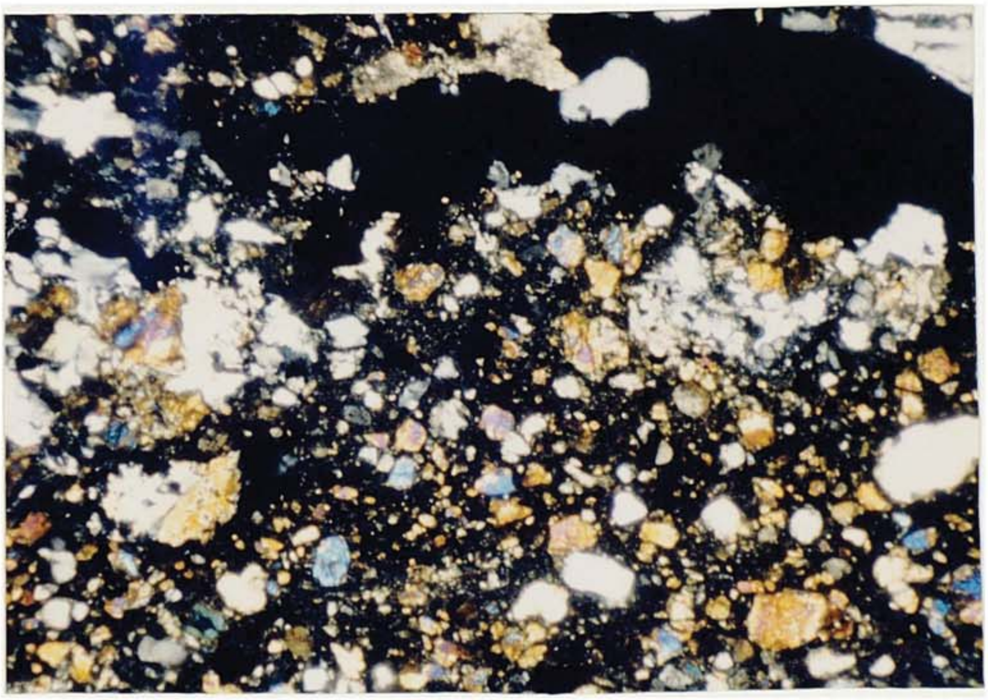


sediments are invariably recrystallised to quartz, albite, minor clinozoisite and prehnite; Fe-rich Ca-Al silicates generally being absent. Calcite is locally abundant, although its presence appears to suppress the development of other phases. Typical assembalges may be comprised of the following;

Silicic volcanics; albite + quartz + chlorite + white mica + sphene ± prehnite ± epidote ± K-feldspar

Silicic volcanoclastics; albite + quartz + chlorite + white mica ± prehnite ± epidote (clinozoisite)

Basaltic volcanoclastics; albite + quartz + chlorite + epidote ± pumpellyite ± prehnite ± stilpnomelane

Whilst not proven beyond doubt, the general absence of Fe-rich silicates in felsic rock types and alternating assamblages within interbedded silicic and basic sediments, strongly suggests a host rock compositional control on phase development. Subtle intergranular variation appears also to have been important, as indicated by highly sericitised feldspar in contact with feldspars which show no signs of alteration other than albitisation. The widespread occurrence of secondary minerals in pelitic and volcanoclastic sediments, indicates that alteration of the coherent volcanics and intrusives is a pervasive event rather than 'spilitisation' associated with hydrothermal alteration as is common to mid-oceanic basalts (e.g. Mevel 1981); although as outlined {5.3.} whilst discussing pépéritic breccias and sediment/magma mixtures, such a process may have been locally operative.

Fault rock mineralogy: Phase development in fault zones and tectono-metadomains also shows an apparent affinity to control by host-rock composition, as may be expected in areas of high fluid-rock ratios. Sedimentary hosted fault rocks commonly contain prehnite, clinozoisite, chlorite, white-mica, and albite (as neocrysts). All of these are enriched in aluminum and relatively depleted in total iron by comparison to basic hosted fault rocks which are dominated by piscasitic epidote, Fe-pumpellyite, prehnite, sphene, stilpnomelane, chlorite, and actinolite. Felsic rock types contain similar assemblages to those of sedimentary fault breccias, although clinozoisite has not been observed. Of greater significance, however, is not the phase assemblages in such rock types, it is their presence alone which indicates that tectonism and metamorphism were coeval at the time of deformation {6.6.}. Certainly there is evidence to indicate that locally phase development would not have arisen unless tectonism had taken place; although this is not to say that metamorphism on a regional scale has arisen directly as a result of tectonism, merely that tectonism may have enhanced a pre-existing metamorphic regime by creating a much higher level of permeability essentially becoming a *fluid-dominated* (Fyfe *et al.* 1978) system, creating open conduits through which the crystallising fluids could flow and neocrysts develop (see section 6.6.).

6.4. MINERAL CHEMISTRY AND FORM

This following section is concerned with a systematic description of primary relict and secondary mineral phases developed in metabasites. The peripheral nature of the primary phase precludes detailed discussion. Where appropriate the chemistry of secondary phases from other lithologies is discussed. Chemical analysis were performed at the Department of Geology, University of Manchester, using a modified Cambridge Instruments Geoscan Mk. II, fitted with a Link System model 290 energy-dispersive spectrometer and ZAF-4/FLS quantitative analysis software system (see Collier 1986). An accelerating voltage of 20Kv was used. Representative analysis are presented Tables (6.1. to 6.11.), which are incorporated in the following text. A more complete list which is fully referenced with specimen numbers, has been deposited with the Departmental Superintendant of the School of Earth Sciences, Birmingham University, Birmingham. Normalizing calculations for Fe^{2+}:Fe^{3+} ratios in amphiboles were carried out with the aid of a computer programme kindly supplied by P.J. Henney, British Geological Survey, Geochemical Division, London.

6.4.1. RELICT IGNEOUS PHASES

(a) Clinopyroxene

Clinopyroxenes have been analysed from gabbros, dolerites, and pillow lavas. Porphyroclastic clinopyroxenes from a shear-zone (Plate 6.4.) have also been analysed (Table 6.1.). Analysis are recalculated on the basis of 6 oxygens, with the assumed formula; $X_{1-p}Y_{1+p}Z_2\,O_6$ where X= Ca and Na, Y= Mg, Fe^{2+}, Fe^{3+}, Al^{vi} and Ti, Z= Si and Al^{iv}, and P= 0-1 (Deer *et al.* 1962). Analysis are plotted on a Di-Hd-En-Fs compositional diagram (Fig. 6.2.a.).

The majority of analysed clinopyroxenes are augitic in composition (Fig. 6.2.a.). In keeping with the findings of Rowbotham & Bevins (1978) and Bevins (1982), who discuss in detail clinopyroxene compositions from various lithologies within the Fishguard Volcanic Complex, clinopyroxenes from extrusives (i.e. pillow lavas) are enriched with regard to their total TiO_2 and Al_2O_3 wt.% oxides (Fig. 6.2.b.) in comparison with clinopyroxenes from intrusives (i.e. dolerites and gabbros). Bevins (1982) attributes this chemical variation to a range in cooling histories as described by Cosley & Taylor (1979) from DSDP basalts, rather than genetically related to melt composition.

Clinopyroxene appears to have been relatively inert to the metamorphic conditions imposed ; an inference substantiated by its relative freshness when observed in thin-section (Plate 6.3.). In all lithologies (excluding metadomains), alteration of clinopyroxene is restricted to chloritisation and more rarely epidotisation along fractures and cleavage planes. Epitaxial amphibole overgrowths, commonly uralitic (Plate 6.3.), can be observed in greenschist assemblages.

Within epidote-metadomains, clinopyroxene is invariably replaced by epidote and less commonly pseudomorphous actinolite, although it may survive as relict variegated patches; suggesting that even under adverse conditions of high-fluid rock ratios the Ca-Mg-Fe-Si-Ti tied up in the structure is not necessarily released. The relatively inert nature of clinopyroxene is also indicated by its persistent presence as porphyroclasts within shear zones and tectono-metadomains; for example, porphyroclastic clinopyroxenes from shear zones cross-cutting dolerite on the Carreg Gybi headland, whilst showing extensive grain size reduction and rounding (Plate 6.4.), exhibit little evidence for replacement or significant compositional variation to clinopyroxenes from elsewhere in the intrusive. The shear zone matrix supporting the porphyroclasts is however, composed of fine acicular pumpellyite and coarse radiate prehnite, presumably forming for the most part after the brakedown and dissolution of feldspar and ore.

(b). Plagioclase feldspar

Primary igneous feldspars, detrital feldspars from the coarse clastic sediments of the Lower Town Formation {4.7.}, and neocrystic plagioclase feldspars from cataclastites, have been analysed (Table 6.2.). Recalculation is based on 8 oxygens in the ternary system $NaAlSi_3O_8$ - $KAlSi_3O_8$ - $CaAl_2Si_2O_8$.

The dominant metabasite feldspar is albite, which is assumed to be reconstituted after a primary precursor calcic-plagioclase of probable labradorite composition. Albitisation is variable however, and a range in composition is observed (Ab_{100} - $Ab_{75}An_{27}Or_2$). This variation is not as extreme as that observed by Bevins (1982; i.e. $Ab_{99}An_1$ - $Ab_{62}An_{38}$), although comparable in the sense that albitisation rarely goes to completion. In all feldspars analysed the maximum anorthite component recorded was 27, whilst rarely does it exceed 10 mole% (Fig. 6.2.c.). No systematic variation in the anorthite component was observed on an intergranular scale, although considerable intra-granular variation is noted on the scale of a thin-section (Ab_{84}-An_{16} to Ab_{94}-An_6). The orthoclase component is recorded to a maximum of 4 mole%, although is commonly absent (Fig. 6.2.c.). Fe, Mg and Ti are detected in small but sporadic amounts. No obvious disparity exists between amphibole bearing and prehnite-pumpellyite facies assemblages. There is no evidence for a peristerite gap in greenschist assemblages the presence of which may indicate an approach to the amphibolite facies (Maruyama *et al.* 1983).

Albite is commonly replaced by other secondary phases, chiefly; pumpellyite, prehnite, chlorite, calcite, white-mica, and to a lesser extent epidote, one or more of which may coexist within a single crystal. Coupled with the presence of polymineralic vesicules and fractures, such replacement indicates at least microdomain scale mobility of the majority of major

Table 6.1. Representative Clinopyroxene analyses from the metabasites within the Fishguard Volcanic Group.

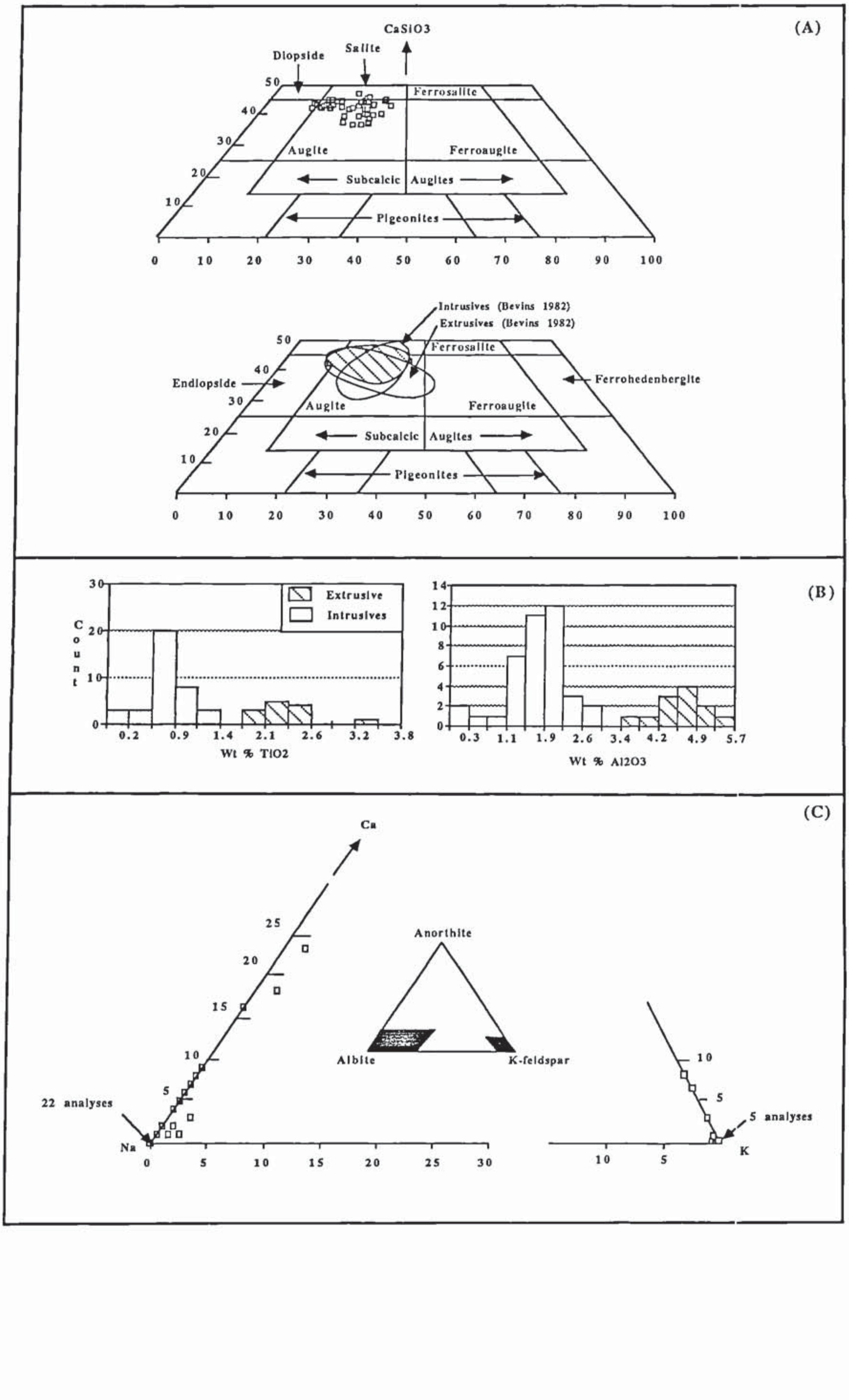
	I	I	I	I	I	I	I	I	I	I	C	C	C	C	E	E	E	E
SiO_2	51.54	53.33	52.16	52.64	52.06	52.74	51.34	50.65	52.39	50.99	50.49	48 26	52.63	51.55	48 27	47 79	47 70	46.01
TiO_2	0.71	0.75	0.74	0.80	0.71	0.58	0.86	1.28	0.57	1.03	1.22	1.80	0.66	0 69	2.28	2.54	2.19	2.61
$Al2O_3$	1.32	1.76	2.28	2.23	1.85	1.94	1.16	1.60	1.82	1.32	1.36	4.91	1.59	1 98	4 77	4.68	4.05	4 99
FeO	12.69	10.27	7.70	7.99	8.29	8.19	14.31	12.93	6.42	14.37	14.25	11.11	7.13	7 48	11.50	11 66	13 07	13.81
MnO	0.44	0.47	0.36	n.d.	0.24	n.d.	0.48	0.59	0.23	0.42	0.45	0.38	n.d.	n.d.	n.d.	0 34	0 44	n.d.
MgO	12.63	12.47	15.42	15.33	15.14	15.38	13.52	13.01	16.98	13.64	12.90	13.01	16.48	16 16	12.40	12.09	12.16	13 81
CaO	19.45	22.25	20.94	21.52	20.85	20.87	18.67	20.19	21.64	18.00	18.95	20.11	20.81	21 07	20.85	21 32	20 33	20.55
Na_2O	0.73	0.55	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.40	n.d.	n.d.	0.52	n.d.	0.47	n.d.	0.46	n.d.	0.72
K_2O	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Total	99.51	101.85	99.69	100.51	99.14	99.71	100.34	100.25	100.45	99.77	99.62	100.10	99.30	99.40	100.07	100.66	99 94	101.35
Recalculation on the basis of 6 oxygens																		
Si	1.959	1.938	1.935	1.927	1.946	1.954	1.942	1.918	1.925	1.933	1.929	1.822	1.950	1.921	1.829	1.803	1 822	1.759
Aliv	0.041	0.062	0.065	0.063	0.054	0.046	0.052	0.072	0.075	0.059	0.061	0.178	0.050	0.079	0 171	0.197	0 178	0 225
Σ	2.000	2.000	2.000	2.000	2.000	2.000	1.994	1.990	2.000	1.992	2.000	2.000	2.000	2.000	2.000	2.000	2.000	1.984
Alvi	0.018	0.016	0.035	0.034	0.028	0.039	0.000	0.000	0.004	0.000	0.000	0.040	0.020	0.008	0.029	0 013	0 005	0.000
Ti	0.020	0.021	0.021	0.022	0.020	0.016	0.025	0.036	0.016	0.030	0.035	0.051	0.018	0.019	0.065	0 072	0 063	0.109
Fe	0.404	0.324	0.241	0.246	0.259	0.254	0.453	0.410	0.197	0.459	0.455	0.351	0.221	0.235	0.365	0.368	0.692	0.442
Mg	0.716	0.702	0.853	0.841	0.844	0.849	0.762	0.753	0.930	0.777	0.735	0.732	0.910	0.898	0.701	0 680	0.692	0.614
Mn	0.014	0.015	0.011	0.000	0.008	0.000	0.016	0.019	0.007	0.014	0.015	0.012	0.000	0.000	0 000	0 011	0 014	0 000
Ca	0.792	0.900	0.823	0.849	0.835	0.829	0.757	0.819	0.852	0.737	0.776	0.813	0.826	0.841	0 846	0 862	0 832	0.842
Na	0.054	0.040	0.000	0.000	0.000	0.000	0.000	0.000	0.029	0.000	0.000	0.030	0.000	0.034	0.000	0.038	0 000	0.054
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000

Table 6.2.) Representative analyses of feldspar from the Fishguard District.

	M	C	C	M	M	M	M	S	S	S	K	K	K	K	K	K	K	K
SiO_2	65.67	66.05	69.35	68.94	68.56	68.86	67.75	70.40	68.65	67.64	63.41	63.06	63.98	64.75	65 18	64 16	65 25	65.38
TiO_2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	2.48	2.22	0.61	0.47	0.46	0.73	0.41	0.31
Al_2O_3	22.13	20.86	19.88	19.72	19.51	18.57	19.05	19.19	18.97	19.95	16.86	16 95	18 24	18.07	18.51	18.39	17 75	17.74
Fe_2O_3	0.52	0.27	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.39	0.91	0.38	0.38	0.45	0.90	n.d.	n.d.
MnO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
MgO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
CaO	3.17	1.87	n.d.	n.d.	0.37	n.d.	0.87	0.20	0.12	1.29	1.49	1.38	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Na_2O	9.41	10.52	11.74	11.42	11.38	10.63	10.41	10.98	10.62	10.30	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.27	n.d.
K_2O	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	14.84	15.02	15 74	15.79	15 99	15.45	15.52	15.71
Total	100.19	99.51	100.97	100.08	99.82	98.06	98.08	100.77	98.36	99.18	99.47	99.54	98.90	99 46	100.59	99.63	99.20	99.14
Structural Formulae on the basis of 8 oxygens.																		
Si	2.856	2.912	3.006	3.001	2.996	3.046	3.004	3.035	3.029	2.974	2.945	2.945	2.984	3 002	2.990	2.974	3.023	3 030
Al	1.145	1.085	1.001	1.001	1.005	0.968	0.996	0.975	0.980	1.034	0.923	0.933	1.003	0.988	1.001	1.005	0.969	0.970
Ti	0.019	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.087	0.078	0.022	0.016	0.016	0.026	0.014	0.011
Fe^{2+}	0.000	0.010	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.038	0.036	0.015	0.015	0.017	0.035	0.000	0.000
Mn	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Mg	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Ca	0.147	0.088	0.000	0.000	0.018	0.000	0.041	0.009	0.006	0.061	0.074	0.069	0.000	0.000	0.000	0.000	0.000	0 000
Na	0.795	0.900	0.973	0.954	0.965	0.912	0.895	0.918	0.911	0.881	0.000	0.000	0.000	0.000	0.000	0.000	0.025	0.000
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.879	0.895	0.937	0.934	0.936	0.914	0.918	0.929
Ab.	84	91	100	100	98	100	96	99	99	93	0	0	0	0	0	0	0	0
An.	16	9	0	0	2	0	4	1	1	7	8	7	0	0	0	0	3	0
Or.	0	0	0	0	0	0	0	0	0	0	92	93	100	100	100	100	97	100

Table 6.1. Representative clinopyroxene analysis form metabasites within the Fishguard Volcanic Complex. (abbreviations, I = intrusive, E= extrusive, C= porphyroclastic clinopyroxenes from a shear on the Carreg Gybi headland). **Table 6.2.** Representative feldspar compositions from a variety of lithologies throughout the Fishguard district (abbreviations, M = metabasites, S = sediments, C = cataclastites, A = silicic volcanics). Also included are several K-feldspar analysis from a basaltic hyaloclastite

Figure 6.2. *Clinopyroxene and feldspar compositional diagrams.* **(Fig. A)** Clinopyroxene compositions plotted in the Di-Hd-En-Fs quadrilateral and comparative diagram with clinopyroxenes analysed by Bevins (1982) from the Fishguard Volcanic Complex. **(Fig. B)** Frequency diagrams of TiO_2 and Al_2O_3 wt.% oxides within clinopyroxenes, showing the relative enrichment of these elements in clinopyroxenes from extrusive (i.e. pillow lavas) and intrusive lithologies (i.e. gabbros and dolerites)., attributed by Bevins (1982) to a range in cooling histories. **(Fig. C)** Feldspar compositions from the Fishguard district plotted on a Ab-Or–An diagram. K-feldspar compositions from a pillow lava and tonalite intrusive.


(A)
CaSiO3
Diopside
Salite
Ferrosalite
Augite
Ferroaugite
Subcalcic Augites
Pigeonites
Intrusives (Bevins 1982)
Extrusives (Bevins 1982)
Endiopside
Ferrohedenbergite
(B)
Extrusive
Intrusives
Count
Wt % TiO2
Wt % Al2O3
(C)
Ca
Anorthite
Albite
K-feldspar
22 analyses
5 analyses
Na
K

elements and possibly some trace elements within a fluid phase. For example, one has to assume that Fe* was introduced from an external source (possibly from the brakedown of ore) to account for phases such as pumpellyite and epidote to be present within the feldspars, whilst in the case of white-mica, Ca and Al will be liberated form the feldspar site with the probable addition of K.

Albite also occurs in a variety of tectonised rocks, crystallising directly from the fluid phase as neocrysts independent of a precursor feldspar. Analysis of neocrystic feldspar (Table 6.2.) indicates a pure end-member composition. Its presence in veins indicates very low μCO_2 in the fluid phase; whilst repeated albite crystallisation in polyphase cross-cutting fractures indicates that at least locally, μCO_2 remained low in time, whilst a_{Na+}, a_{Al+} and a_{Si+} remained high.

The general scarcity of K-feldspar and the predominance of albite in rhyolites may indicate rock-water temperatures well in access of 140°C, as lower temperatures generally favour adularia in felsic rock types (Munhá *et al.* 1980), although albite predominance may also reflect high a_{Na+}/a_{K+} values (Brown & Ellis 1970). Feldspar compositions compare favourably with metabasites from elsewhere in Wales (see Bevins & Rowbotham 1983), the presence of a small, though variable, anorthite component being typical of low grades of metamorphism.

(c). Ore (Fe–Ti oxides)

Ore is ubiquitous to all metabasites and invariably shows evidence for at least partial breakdown. It is common in coarser grained intrusives to observe the ilmenite fraction pseudomorphed by sphene, leaving a mosaic magnetite fraction as a relict oxide. In finer grained extrusives the opaque mineralogy is often difficult to discern, invariably the result of replacement by dusty aggregates of sphene. It has been shown that iron-enrichment of Ca-Al silicates owes much to the brakedown of opaques (Offler *et al.* 1981). It seems likely that the brakedown of ore was a contributory factor, in conjunction with the physical parameters of metamorphism, in producing the widespread development of Fe-rich Ca-Al silicates.

6.4.2. SECONDARY MINERALS

(a). Chlorite

Chlorite is pervasive in metabasites throughout the area. Its form is variable although characteristically occurs as;

a) laths and plates interstitial to the primary mineralogy
b) a replacement of relict phases, commonly along structural weakness
c) a monomineralic, or a component of polymineralic vesicular and vein assemblage, where it may exhibit coarse radiate and colloform textures.
d) a psuedomorphic phase of pillow lava margins and hyaloclastites

Representative chlorite analyses are presented in Table (6.3.). Recalculation is based on 28 oxygens, assuming the general formula; $Y_{12}Z_8O_{20}(OH)_{16}$, where Z = Si and Al^{iv} and Y = Al^{vi}, Fe^{3+}, Fe^{2+}, Mn, Mg, and Ca, Na, K. The recalculation assumes the tetrahedral sites are filled with Si and Al to a maximum of 8, with residual Al as Al^{vi} partitioned into the octahedral sites. Fe is computed as FeO.

The main chemical features of the metabasite chlorites are; tetrahedrally co-ordinated Al varies from 1.53 to 2.01 reciprocal to Si, whilst Y-site occupancy varies from 11.43 to 12.02. Al^{vi} is variable from 1.98 to 4.78 and exceeds Al^{iv} in all cases; although this apparent charge imbalance appears to be common to chlorites from low-grade terrains (e.g. Viereck *et al* 1982). The ratio Fe*/Fe* + Mg is much larger than the limited values for the ratio Al*/Fe* + Mg, suggesting iron is largely in the ferrous state. The presence of small amounts of CaO, Na_2O and K_2O, with cation totals invariably less than 20 (see, Foster 1962 for vacancies in the chlorite structure) may be related to interlayered clays, the presence of minor alkalis being typical of chlorites from low grade terrains. To reveal chemical heterogeneity, analyses are plotted on an Si *vs*. Fe* compositional diagram (Fig. 6.3.) of Hey (1954). According to this widely used classification, the dominant metabasite chlorites are pychnochlores and brunsvigites, with limited inter-specimen variation suggestive of an approach to equilibrium. Sedimentary chlorites are more variable with regards their compositional range, spanning the ripidolite, pychnochlorite, brunsvigite and diabonite fields, whilst alkali contents are appreciably higher.

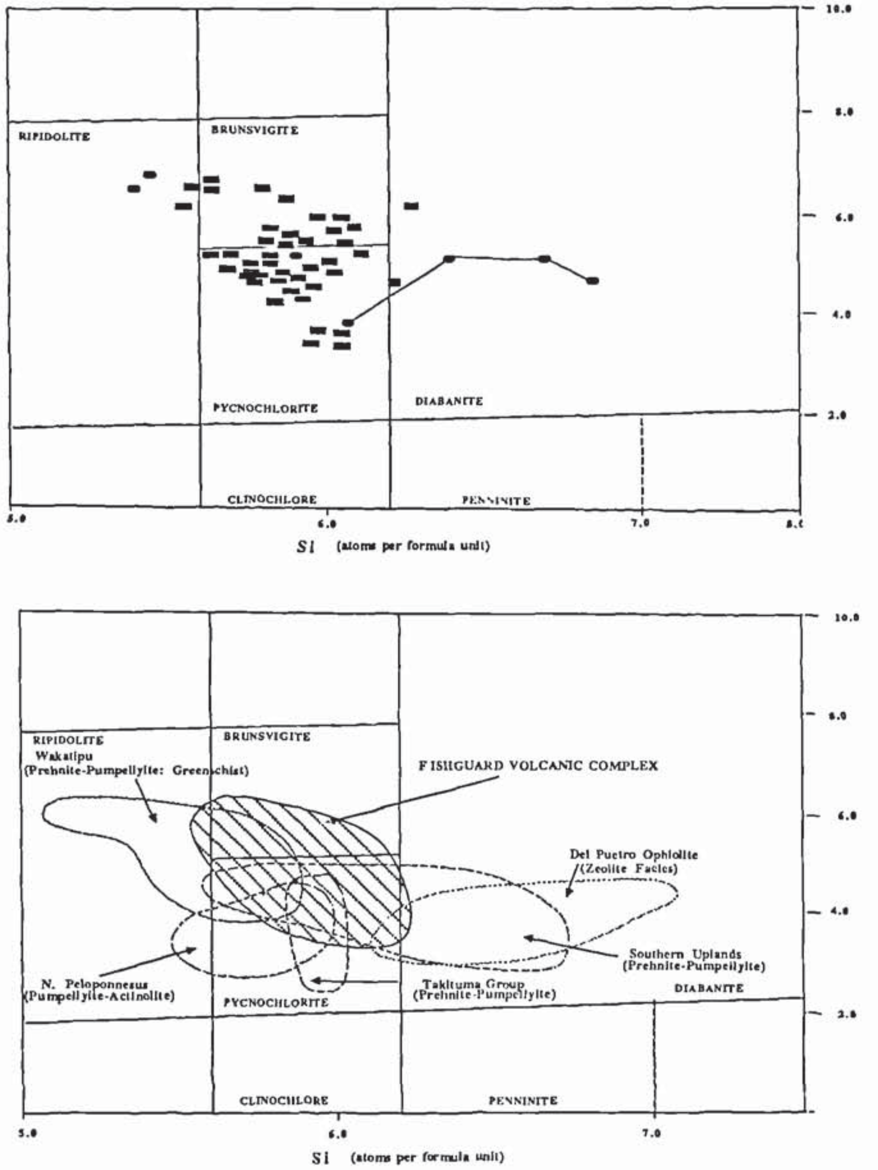
It has been demonstrated (e.g. Maruyama *et al.* 1983, Cho *et al.* 1986) that subtle compositional change within chlorite may occur in response to varying metamorphic conditions; particularly temperature and specifically within areas of high geothermal gradients (Fig. 6.3.b.). For example, it has been shown that with increasing metamorphism, a decrease in the ratio X_{Fe*} $(=n_{Fe*}/n_{Fe*} + n_{Mg*})$ is common, as is an increase in the substitution of amesite (Al + Al) for antigorite ([Mg, Fe*] + Si). Whilst the composition of chlorites observed in the Fishguard metabasites compare favourably with those from other prehnite-pumpellyite and greenschist facies terrains (Fig. 6.3.a.) and are significantly more silicic than those from zeolitic terrains, no detectable trend in chlorite composition from the various assemblage types is apparent (Fig. 6.3.b.). This may suggest the influence of host rock chemistry in determining chlorite composition (Bevins & Rowbotham 1983) may have masked any change with increasing grade, although a more detailed chemical survey may as yet reveal subtle changes, as a high geothermal gradient to metamorphism is implied {6.6.}.

(b). Sphene

Sphene is ubiquitous throughout the metabasites of the district. It is common in intermediate

Figure 6.3. *Chlorite compositional diagrams.* **(Fig. A).** Chlorite compositional plots of Si vs. Fe* on Hey's diagram (Hey 1954) with comparative diagram including compositional fields from other low-grade terrains (Wakatpui – Kawachi 1974; N. Peloponnesus – Katagas & Panagos1984; Del Puerto – Evarts & Schiffman 1983; Southern Uplands – Oliver & Legget 1980; Takituma Group – Houghton 1982). **(Fig. B).** SiO_2 *vs.* Al_2O_3 plot for compositions of metabasite chlorites from the Fishguard district compared with those of Maruyama *et al.* 1983 and others from various low grade terrains within the Karmutsen area of British Columbia.

(A)


RIPIDOLITE
BRUNSVIGITE
PYCNOCHLORITE
DIABANITE
CLINOCHLORE
PENNINITE
Total Fe
Si (atoms per formula unit)
Wakatipu
(Prehnite-Pumpellyite: Greenschist)
FISHGUARD VOLCANIC COMPLEX
Del Puetro Ophiolite
(Zeolite Facies)
Southern Uplands
(Prehnite-Pumpellyite)
N. Peloponnesus
(Pumpellyite-Actinolite)
Takituma Group
(Prehnite-Pumpellyite)

(B)

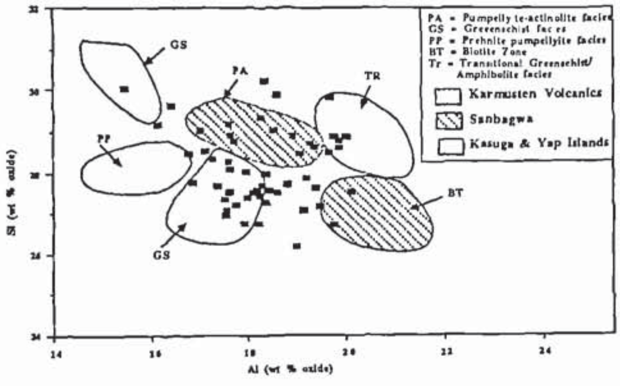

PA = Pumpellyite-actinolite facies
GS = Greenschist facies
PP = Prehnite pumpellyite facies
BT = Biotite Zone
Tr = Transitional Greenschist/Amphibolite facies
Karmusten Volcanics
Sanbagwa
Kasuga & Yap Islands
GS
PA
TR
PP
BT
Si (wt % oxide)
Al (wt % oxide)

Table 6.3. Chlorite analyses from the Fishguard District.

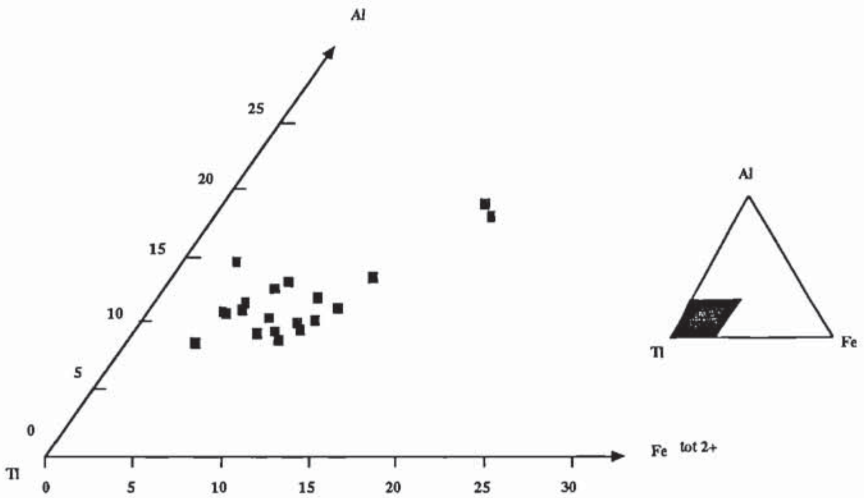
	C	C	C	M	M	M	M	M	M	M	M	M	M	S	S	S	S
SiO2	25.84	27.50	27.58	26.73	25.69	27.59	27.85	28.89	30.68	26.08	28.80	28.61	27.66	25.20	25.99	26.73	27.16
TiO2	n.d	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Al2O3	18.19	17.59	18.44	18.19	22.99	18.58	19.20	20.01	14.17	20.19	19.85	19.35	17.34	22.38	23.30	17.92	19.43
FeO*	33.52	32.43	27.37	30.72	32.65	27.06	26 81	24.11	27.20	34.39	26.03	21.94	29.31	30.98	31.13	31.75	29.18
MnO	0.43	n.d.	0.32	0.69	n.d.	0.32	0.43	n.d.	n.d.	0.76	0.31	n.d.	0.55	n.d.	n.d.	0.45	0.60
MgO	10.48	11.10	14.35	12.72	9.29	14.53	14.73	17.30	15.48	7.91	14.07	17.19	13.78	9.19	9.24	11.02	13.79
CaO	n.d.	0.15	n.d.	n.d.	n.d.	n.d.	0.20	0.19	0.56	n.d.	n.d.	0.14	n.d.	n.d.	n.d.	0.22	0.39
Na2O	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
K2O	n.d.	n.d.	0.19	n.d.	n.d.	n.d.	0.18	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.11	n.d.	n.d.	n.d
Total	88.46	88.77	88.26	89.05	90.62	88.08	89.40	90.50	88.09	89.33	89.06	87.37	88.64	87.86	87.66	88.09	90.55
Si	5.659	5.926	5.912	5.831	5.402	5.827	5.788	5.806	6.456	5.650	5.931	5.900	5.889	5.443	5.472	5.758	5.605
Aliv	2.341	2.147	2.169	2.280	2.592	2.173	2.212	2.194	1.535	2.350	2.069	2.100	2.111	2.557	2.528	2.242	2.395
ΣZ	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Alvi	2.353	2.293	2.427	2.307	3.147	2.452	2.490	2.546	1.984	2.807	2.750	2.650	2.242	2.139	2.254	2.307	2.399
Ti	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Fe2+	6.138	5.845	4.840	5.496	5.739	4.778	4.660	4.053	4.793	6.429	4.483	3.784	5.219	5.595	5.483	5.719	3.573
Mn	0.080	0.000	0.057	0.126	0.000	0.059	0.076	0.000	0.000	0.141	0.054	0.000	0.099	0.000	0.000	0.076	0.102
Mg	3.421	3.565	4.532	4.058	2.918	4.574	4.565	5.183	4.863	2.556	4.321	5.285	4.375	2.960	2.901	3.706	4.388
Ca	0.000	0.034	0.000	0.000	0.000	0.000	0.045	0.041	0.135	0.000	0.000	0.032	0.000	0.000	0.000	0.052	0.084
Na	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
K	0.000	0.000	0.051	0.000	0.000	0.000	0.049	0.000	0.000	0.000	0.000	0.000	0.000	0.030	0.000	0.000	0.000
ΣCat	19.99	19.74	19.90	19.98	19.81	19.86	19.88	19.82	19.79	19.89	19.71	19.79	19.94	19.72	19.74	19.86	19.99
XFe	0.64	0.62	0.52	0.58	0.66	0.51	0.50	0.43	0.49	0.71	0.51	0.41	0.55	0.65	0.65	0.60	0.53

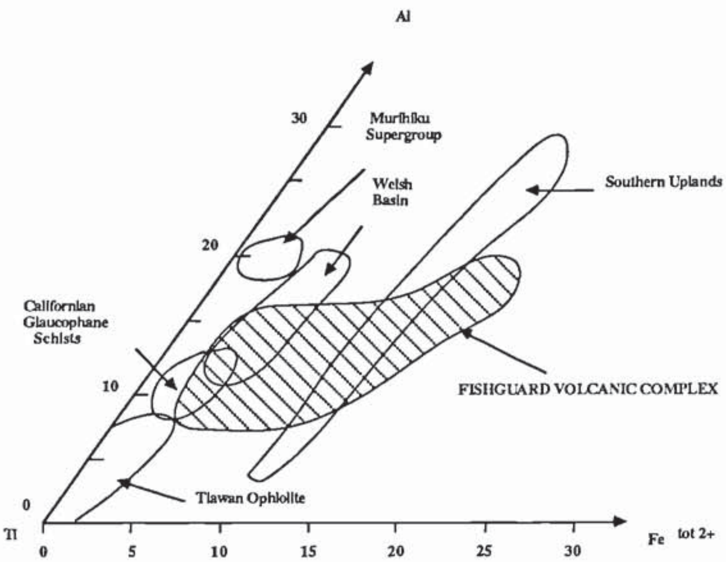
Table 6.4. Representative analyses of sphene

SiO_2	31.76	31.46	31.66	31.21	31.08	31.16	30.88	29.66	30.75	31.08	30.53	30.79	30.77	30.49	30.25	30.80	31.04
TiO_2	33.39	27.61	34.17	36.06	29.99	34.18	33.99	35.47	31.46	36.34	36.93	31.06	36.92	34.93	34.47	34.55	35.83
Al_2O_3	3.45	5.60	3.59	2.64	5.05	2.79	2.66	3.46	3.62	3.36	3.17	5.42	3.10	3.02	2.73	2.87	4.11
FeO*	2.36	7.81	2.59	1.43	6.13	2.57	2.67	3.59	4.30	2.07	1.96	6.07	1.59	1.72	3.11	3.46	1.18
MnO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
MgO	0.32	3.54	0.45	0.26	2.29	n.d.	n.d.	0.63	1.43	0.58	0.30	2.42	n.d.	0.39	n.d.	n.d.	0.38
CaO	28.95	21.96	28.22	28.05	24.10	28.70	28.19	27.01	26.05	26.32	26.84	22.73	27.42	27.38	27.79	27.51	27.92
Na_2O	n.d.	0.28	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
K_2O	n.d.	0.36	0.14	n.d.	0.20	n.d.	n.d.	n.d.	0.25	n.d.	0.18	0.26	0.12	0.13	0.15	n.d.	0.21
Total	100.23	98.62	99.82	99.65	98.84	99.40	98.39	99.82	97.87	99.75	99.91	99.48	99.92	98.06	98.50	99.19	100.67
Recalculation based on 4 Si cations																	
Si	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000
Ti	3.254	2.639	3.244	3.665	2.907	3.299	3.410	3.594	3.070	3.613	3.638	3.034	3.606	3.647	3.626	3.469	3.471
Al	0.511	0.840	0.536	0.370	0.823	0.422	0.407	0.550	0.555	0.507	0.491	0.832	0.473	0.408	0.423	0.440	0.625
Fe^{3+}	0.235	0.521	0.222	0.000	0.261	0.275	0.193	0.000	0.375	0.000	0.000	0.134	0.000	0.000	0.000	0.087	0.000
Σ Y	4.000	4.000	4.000	4.035	4.000	3.996	4.000	4.144	4.000	4.120	4.129	4.000	4.079	4.150	4.049	4.000	4.096
Ca	3.907	2.996	3.834	3.910	3.323	3.957	4.035	3.900	3.630	3.768	3.768	3.162	3.815	3.855	3.934	3.826	3.885
Mg	0.058	0.672	0.083	0.050	0.442	0.000	0.000	0.000	0.277	0.111	0.060	0.468	0.000	0.077	0.000	0.000	0.074
Na	0.000	0.071	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.082	0.000	0.000	0.000
K	0.000	0.052	0.024	0.000	0.032	0.000	0.000	0.000	0.041	0.000	0.003	0.044	0.019	0.022	0.025	0.000	0.034
Fe^{2+}	0.013	0.309	0.053	0.155	0.402	0.000	0.192	0.404	0.096	0.221	0.216	0.595	0.174	0.189	0.342	0.289	0.128
Σ X	3.978	4.100	4.004	4.115	4.119	3.957	4.254	4.304	4.004	4.100	4.047	4.269	3.998	4.230	4.301	4.115	4.121

Table 6.3. Representative chlorite analysis from the Fishguard district (abbreviations M = metabasite chlorites, C = cataclastite chlorites, S = sedimentary chlorites). **Table 6.4.** Representative sphene analysis from metabasites within the Fishguard Volcanic Complex.

Fig. 6.4. Al-Ti-Fe diagram showing compositional variation in analysed sphene from the Fishguard metabasites. Comparative plot includes sphenes from a variety of terrains (see Veireck *et al.* 1982, Oliver & Legget 1980, Bevins & Rowbotham 1983).


Al
25
20
15
10
5
0
Ti
0
5
10
15
20
25
30
Fe tot 2+
Al
Ti
Fe


Al
30
20
10
0
Ti
0
5
10
15
20
25
30
Fe tot 2+
Murihiku
Supergroup
Welsh
Basin
Southern Uplands
Californian
Glaucophane
Schists
FISHGUARD VOLCANIC COMPLEX
Tiawan Ophiolite

lithologies, and may be observed in rhyolitic volcanics as an accessory phase. It is most commonly observed as fine frambiodal aggregates in which individual crystals are difficult to distinguish (Plate 6.1.), although readily identifiable due to its very high relief and extreme birefingence; weak pleiochroism may be evident in darker coloured varieties. More rarely it is observed as discrete tabular crystals, a form common to sphenes hosted within silicic volcanics.

Sphene analyses from several basic lithologies are presented in Table (6.4.), wholly reliable analyses were difficult to obtain due to its fine grained frambiodal nature, although cation totals are relatively constant and an appreciation of the chemistry can be obtained. Recalculation is based on the assumption of 4 Si cations in the unit cell, according to the formula $X_4Y_4Si_4(O,OH,F)_{20}$; assuming Si fills the tetrahedral sites, and after assigning Ti, Al and Fe^{3+} to fill the Y-sites, excess Fe as Fe^{2+} along with Ca, Mg, Na, and K are placed in the X-sites (*c.f.* Coombs *et. al.* 1976), although noting there is no independent analytical evidence for Fe^{2+} in the sphene structure (Higgins & Rubbie 1976).

The general chemical characteristics of the sphenes analysed, indicate both a considerable variation in TiO_2 content (26.6 to 38.0 wt. % oxide) and grothite component (Ca [Al, Fe] SiO_4 (OH)). SiO_2 and CaO are relatively constant with Y-site occupancy close to 4, whilst X-site occupancy is invariably greater than 4; primarily the result of partitioning excess Fe into this site. The dominant substitution appears to be Al+Fe^{3+} for Ti to a maximum of 1.36 cations per formula unit. Whilst this substitution is governed by complex kinetic factors it has been suggested as typical of sphenes from low-grade terrains (Boles & Coombs 1977). MgO is detected in small sporadic amounts, as is Na_2O and K_2O, the significance of which is unclear, given that analysis are not wholly reliable. However, sphene compositions compare favourably with those from elsewhere in Welsh Basin (Bevins & Rowbotham 1983) and those recorded by Moore-Biot (1970) from Permo-Carboniferous spilites (Fig. 6.4.). They are however, distinctly enriched with regards total iron and to a lesser extent aluminum than those recorded from prehnite-actinolite facies (e.g. Liou & Ernst 1979) and blueschist terrains.

Schuiling & Vink (1967) showed that the presence of sphene is favoured by high P_{fliud}, fO_2 and P_{H20}, with the reaction $CaTiSiO_5 + CO_2 <=> CaCO_3 + SiO_2 + TiO_2$ limited to $P_{CO2} < 200$ bars at 450°C and $P_{CO2} = 50$ bars at 350°C; effectively restricting sphene development in the Fishguard metabasites to very low P_{CO2}, probably less than 50 bars. However, Viereck *et al.* (1982) demonstrate, that apart from factors such as temperature and P_{H2O}, precursor mineral chemistry may also be influential in determining sphene composition and paragenesis. Regarding this, it is of interest to note that sphene from felsic volcanics within the district are distinctly lighter in colour, possibly reflecting a lower FeO* content as a likely consequence of lower whole rock total iron.

(c) Pumpellyite

Pumpellyite was first identified in metabasites of the Fishguard district by Bevins (1978), leading him to suggest that the region had been subjected to low-grade metamorphism during Caledonian times. In keeping with the findings of Bevins (1978) pumpellyite has been identified in a variety of forms, including;

a) discrete acicular and rod-like laths and more rarely radiate fans replacing feldspar
b) amorphous cryptocrystalline aggregates within the groundmass,
c) discrete bow-tie and radiate rosettes, common to cataclastites (Plate 3.4.)
d) amorphous to subhedral aggregates in polymineralic vesicular infill

Representative analyses of pumpellyite from a variety of basic rock types, including cataclastites, are presented in Table (6.5.). Recalculation is based on 16 cations per formula unit, according to the formula; $W_4X_2Y_4Z_6\,O_{(20+x')}\,(OH)_{(8-x)}$, where W= Ca, Mn, and K, X= $(Mg, Fe^{2+}, Mn)_{2-x'}\,(Fe^{3+}, Al)_{x'}$, Y= Fe^{3+} and Al, and Z= Si (+Al); *c.f.* Coombs *et al.* (1976). The cations are assigned to the four sites according to the scheme of Passaglia & Gottardi (1973).

The general features of those pumpellyites analysed are; that they classify as Fe-pumpellyites (Fig. 6.5.), being compositional intermediate between the theoretical Al-end member and Fe^{2+}: Fe^{3+} species jugöldite. Al is generally absent from the tetrahedral sites, as Si is generally close to the ideal value of 6. Al in the octahedral sites is close to 4, although Fe (assumed to be Fe^{3+}) is required in small amounts up to a maximum of 0.51 to satisfy a total of 4 Y-site cations. X-site occupancy is invariably close to 2, although minor deficiency is apparent from several analysis; this may not be totally unexpected as it may be incorrect to assume a constant number of divalent cations in the X-site (Coombs *et al.* 1976). Ca in the W-site is generally greater than 4, indicating Fe^{2+}<—> Ca exchange is unlikely. The Fe^{2+}: Fe^{3+} can not be solved as the oxidation state is unknown, although Liou (1979) argues that Fe* in the pumpellyite structure can be assumed to be trivalent due to the common antipathetic relationship of Fe–Al. However, when Fe* *vs.* Al and Fe* *vs.* Al+Mg are plotted (Fig. 6.5.b.), a strong linear correlation is observed in both plots, suggesting that at least some of the total iron present within the pumpellyite structure is present in the ferric state with likely substitutions of Fe^{3+}<—> Al and Fe^{2+}<—> Mg.

To show chemical heterogeneity and inter-specimen variation, analyses are plotted on Al:Fe*:Mg triangular diagrams (Fig. 6.5.a.). It is apparent that whilst a wide range in composition is observed, inter-specimen variation is limited with regards to Mg (<10 mole%) and to a lesser degree Fe*<—>Al. It has been widely recognised that Fe*<—>Al substitution is in part related to metamorphic conditions, in particular an inverse relationship with pressure (e.g. Coombs *et al.* 1976, Trzcienski & Birkett 1982). Regarding this relationship, pumpellyite

Figure 6.5. *Pumpellyite compositional diagrams.* **(A)** Triangular Al-Fe*-Mg plot of pumpellyite compositions from the Fishguard metabasites with comparative plot from various terrains (see Coombs *et al.* 1976, Offler *et al.* 1981, Oliver & Legget 1980). **(B)** Total Fe *vs.* Al+Mg and Total Fe *vs.* Al.

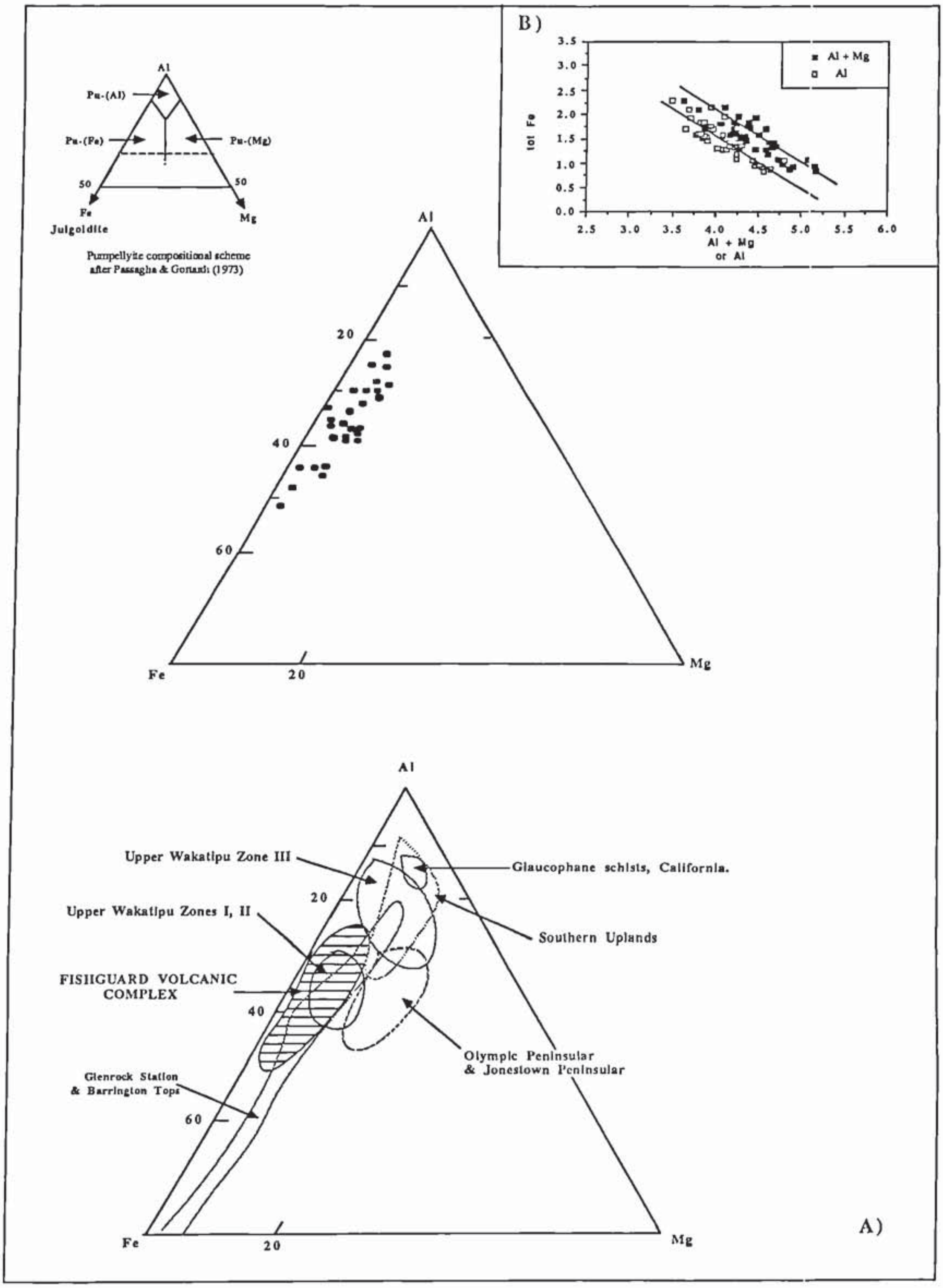

B)
3.5
3.0
2.5
2.0
1.5
1.0
0.5
0.0
tot Fe
2.5
3.0
3.5
4.0
4.5
5.0
5.5
6.0
Al + Mg
or Al
Al + Mg
Al
Al
Pu-(Al)
Pu-(Fe)
Pu-(Mg)
50
50
Fe
Mg
Julgoldite
Pumpellyite compositional scheme
after Passaglia & Gottardi (1973)
Al
20
40
60
Fe
20
Mg
Al
Upper Wakatipu Zone III
Glaucophane schists, California.
20
Upper Wakatipu Zones I, II
Southern Uplands
FISHGUARD VOLCANIC
COMPLEX
40
Olympic Peninsular
& Jonestown Peninsular
Glenrock Station
& Barrington Tops
60
Fe
20
Mg
A)

Table 6.5 Representative analyses of pumpellyite form the Fishguard District.

	M	M	M	M	B	B	M	M	M	M	M	M	C	C	C	C	C	C
SiO_2	36.17	36.18	36.12	36.57	36.89	36.24	36.64	36.61	36.73	37.99	36.56	36.63	37.11	37.19	37.21	37.19	36.76	37.15
TiO_2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Al_2O_3	19.22	19.76	17.70	19.78	21.19	19.74	20.36	21.26	19.83	24.53	22.39	19.90	19.99	19.99	20.28	20.27	19.93	19.68
FeO^*	15.46	13.01	16.41	13.05	11.36	13.32	12.01	11.84	13.43	6.48	9.92	11.04	11.84	11.12	11.95	10.74	11.38	11.69
MnO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
MgO	0.52	1.00	0.55	0.77	1.02	1.41	1.02	0.57	2.29	0.94	1.83	1.69	1.75	178	1.71	1.88	1.76	1.65
CaO	23.34	23.20	22.54	22.92	23.17	22.15	22.78	22.83	22.39	24.15	23.44	23.12	23.69	23.37	23.27	23.46	23.33	23.53
Na_2O	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
K_2O	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Total	94.71	93.15	93.32	93.09	93.63	92.86	92.81	93.11	93.67	94.09	94.14	92.38	94.38	93.45	94.42	93.54	93.16	93.70
Structural formulae on the basis of 16 cations																		
Si	6.032	6.027	6.044	6.062	6.041	6.008	6.058	6.040	6.015	6.097	6.002	6.065	6.023	6.067	5.861	6.069	6.040	6.074
Al^Z	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.139	0.000	0.000	0.000
ΣZ	6.032	6.027	6.044	6.062	6.041	6.008	6.058	6.040	6.015	6.097	6.002	6.065	6.023	6.067	6.000	6.069	6.040	6.074
Al^Y	3.677	3.644	3.487	3.865	4.000	3.858	3.967	3.866	3.828	4.000	4.000	3.885	3.824	3.850	3.626	3.900	3.860	3.792
Ti	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Fe^{3+}	0.323	0.366	0.513	0.135	0.000	0.142	0.033	0.144	0.182	0.000	0.000	0.115	0.176	0.150	0.374	0.100	0.140	0.208
ΣY	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000
Al	0.000	0.000	0.000	0.000	0.089	0.000	0.000	0.000	0.000	0.644	0.219	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Fe^{2+}	1.775	1.335	1.780	1.674	1.555	1.653	1.666	1.660	1.440	0.868	1.326	1.414	1.413	1.369	1.200	1.366	1.424	1.390
Mg	0.125	0.232	0.137	0.191	0.248	0.347	0.251	0.200	0.560	0.227	0.436	0.419	0.424	0.434	0.402	0.457	0.431	0.404
ΣX	1.900	1.567	1.917	1.865	1.892	2.000	1.917	1.860	2.000	1.739	1.981	2.835	1.855	1.803	1.602	1.723	1.856	1.794
Ca	4.058	3.888	4.037	4.071	4.065	3.936	4.035	4.012	3.745	4.152	4.015	4.012	4.100	4.091	3.928	4.105	4.017	4.102
Fe	0.000	0.000	0.000	0.000	0.000	0.052	0.000	0.000	0.222	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
K	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
ΣW	4.058	3.888	4.037	4.071	4.065	3.978	4.035	4.012	3.976	4.152	4.015	4.012	4.100	4.091	3.928	4.105	4.017	4.102

Table 6.6 Representative analyses of prehnite from the Fishguard District (XFe* = 100 Fe/Fe + Al).

	C	C	M	M	M	M	M	M	M	M	C	C	M	S	C	C
SiO_2	44.88	42.28	44.75	43.84	44.69	44.13	42.84	43.49	43.67	42.82	44.29	44.93	43.72	44.00	43.67	43.17
TiO_2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.25	0.26	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Al_2O_3	23.72	22.18	23.19	21.85	23.10	23.41	23.42	23.52	20.29	20.55	23.45	24.48	24.40	23.81	23.43	23.14
Fe_2O_3	0.29	1.35	0.74	2.94	0.84	1.30	3.17	0.57	5.19	5.36	0.28	n.d.	0.54	0.63	0.72	0.76
MnO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
MgO	n.d.	n.d.	n.d.	n.d.	n.d.	0.41	1.99	n.d.	n.d.	0.23	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
CaO	27.61	27.04	28.06	27.94	28.23	27.10	24.29	28.71	27.58	26.57	28.94	28.43	28.66	28.60	27.84	27.63
Na_2O	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
K_2O	n.d.	n.d.	n.d.	n.d.	n.d.	0.18	0.23	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Total	96.50	92.85	96.74	96.04	96.86	96.53	96.19	96.55	96.73	95.53	96.96	97.84	97.33	97.04	95.66	94.70
Structural formulae based on 22 oxygens																
Si	3.086	3.023	3.053	3.050	3.017	3.024	2.959	2.985	3.057	3.033	3.018	3.021	3.004	3.032	3.013	3.032
Al^{iv}	0.914	0.997	0.947	0.950	0.983	0.976	1.041	1.015	0.943	0.967	0.982	0.979	0.996	0.968	0.997	0.968
ΣZ	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000	4.000
Al^{vi}	0.966	0.892	0.918	0.815	0.943	0.914	0.865	0.924	0.731	0.740	0.938	0.961	0.936	0.923	0.919	0.940
Ti	0.000	0.000	0.000	0.000	0.000	0.000	0.013	0.013	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.013
$Fe3+$	0.017	0.080	0.045	0.171	0.048	0075	0.183	0.032	0.304	0.318	0.015	0.000	0.030	0.036	0.042	0.045
ΣY	0.983	0.972	0.963	0.986	0.982	0.989	1.061	0.969	1.035	1.068	0.953	0.961	0.966	0.959	0.975	0.998
Mg	0.000	0.000	0.000	0.000	0.000	0.046	0.205	0.000	0.000	0.024	0.000	0.000	0.000	0.000	0.000	0.000
Ca	1.990	2.071	2.051	2.041	2.042	1.989	1.797	2.064	2.069	2.017	2.066	2.048	2.063	2.065	2.058	2.004
Na	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
K	0.000	0.000	0.000	0.000	0.000	0.016	0.020	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
ΣX	1.990	2.071	2.051	2.041	2.042	2.011	2.023	2.064	2.069	2.041	2.066	2.048	2.063	2.065	2.058	2.004
XFe^*	8.23	1.49	4.67	17.34	4.94	7.58	17.46	3.34	29.37	30.05	1.59	0.00	3.10	3.75	4.37	4.57

Table 6.5. Representative pumpellyite compositions from metabasites (M), basic tuffs (B) and metabasic cataclastites (C) of the Fishguard district. **Table 6.6.** Representative prehnite analysis from metabasites (M) and cataclastites (C) of the Fishguard district.

from the Fishguard Volcanic Complex shows affinity with Fe-rich pumpellyite from documented low pressure terrains (Fig. 6.5.b.), although deficient in Al with respect to those intermediate and high pressure regimes.

Whilst a P_{tot} – Fe* inverse relationship is apparent in many terrains, and is appropriately applied here, it seems unlikely that ardent adherence to this widely observed relationship is wholly appropriate in the light of suggestions that a_{Fe*} in the fluid phase (Offler *et al.* 1981), fO_2 (Nystrom 1983) and whole rock chemistry (Oliver *et al.* 1980), may be equally as important in determining pumpellyite composition. It is of interest to note, that in a recent study by Cho *et al.* (1986) concerning the zeolite to prehnite-pumpellyite facies transition within the Karmutsen Volcanics, British Columbia, the Fe* content of pumpellyite was observed to increase, in effect opposite to the P_{tot} – Fe* inverse relationship cited above. Nevertheless, the high Fe* content of pumpellyite from the Fishguard metabasites is in character with the estimated P–T conditions that prevailed {6.6.}, and whilst a_{Fe*} and fO_2 were possibly high during prograde conditions, low load pressure is likely to have had a major role in determining pumpellyite composition.

(d). Prehnite

Prehnite is widely developed throughout the pillow lavas and high level intrusives on the Pen Caer peninsular. It is generally absent in rocks containing amphiboles, although may be evident in minor amounts within 'transitional' type assemblages {6.2.}. It is a common constituent of basic volcanoclastics, pillow breccias and hyaloclastics, and may locally be observed as an amorphous aggregate within pelitic sediments, often preferentially enhancing diffuse biogenic structures. It is common to acidic volcanics and felsic intrusives on the Pen Caer peninsular, frequently associated with chlorite, sphene, albite, white-mica, ± epidote. It is a major neocrystic component of cataclastites throughout the upper parts of the Strumble Head Volcanic Formation. In thin-section prehnite form is variable, most commonly occurring as;

a) platy laths, exhibiting radiate, spherulitic and more rarely bow-tie textures,
b) amorphous cryptocrystalline aggregates within the groundmass or vesicules,
c) mammillary aggregates with an ameboidal replacement form,
d) rare euhedral crystals,
e) spongy and variegated cryptocrystalline aggregates, showing evidence of breakdown.

In thin-section prehnite is generally colourless to faint light brown, the latter probably reflecting slight iron enrichment. Interference colours are invariably low to mid-second order, wavy extinction may be apparent. Of passing interest, is the ability of prehnite to luminesce yellow (Plate 6.5.a,b). Such a technique highlights crystal growth which is not observed by standard microscopy. It has been found to be a useful petrographic tool in determining the replacement of feldspars by fine prehnite flakes. Whilst requiring further work regarding quantification of luminescence, this property appears not to be documented; it may prove to be a useful source of information in determining semiquantative phase mineralogy in fine grained metabasites. Variation in luminescence may be attributed to Fe-enrichment in the cores of crystals and depletion in rims, attributed to increasing fO_2 during prograde conditions.

Analyses of prehnite from a variety of rock types are presented in Table (6.6.). Recalculation is based on 11 oxygens and all iron is assumed to be Fe_2O_3, the ideal formula being; $X_2YZ_4\,O_{10}\,(OH)_2$, where X= Ca, Mg, K, and Mn; Y= Al^{vi}, Fe^{3+}; Z= Si, Al^{iv} (Surdam 1969).

The chemical characteristics of prehnite indicate that Si is invariably close to the ideal value of 3, with tetrahedrally co-ordinated Al close to 1. Y-site occupancy is close to 1, with $Fe^{3+} \longleftrightarrow Al^{vi}$ substitution in the range $0.00 \longleftrightarrow 0.299$ mole fraction of $Ca_2Fe^{3+}(Al,Si_3)\,O_{10}\,OH_2$, the range of substitution is however, generally limited to less than 0.100. Ca is generally in slight excess of 2, indicating X-site substitution is unlikely. Ti, K, Mg and Mn are present in small sporadic amounts. Intra- and intergranular variation may locally be pronounced, particularly in cataclastite assemblages. The ratio $100(Fe^{3+}/(Fe^{3+} + Al^{tot}))$ has been recorded to vary in the range 1.33 to 11.63 for intergranular composition of prehnite from a cataclastite, whilst values of 1.33 and 5.25 are recorded from the rim and core respectively of a single crystal from the same specimen.

Chemical heterogeneity of analysed prehnite is diagrammatically represented in a Ca:Al:Fe* triangular diagram (Fig. 6.6.). The observed linear relationship reflects the strong tendency for $Fe^{3+} \longleftrightarrow Al$ exchange. Using a comparative triangular diagram (Fig. 6.6.), it is apparent that whilst a limited range in prehnite composition is evident, they compare favourably with metabasites from elsewhere in Wales (Bevins & Rowbotham 1983) and other low-grade terrains (e.g. Kay 1984) although distinctly more aluminous than others (e.g. Kuniyoshi & Liou 1976). It has been suggested that iron-rich prehnite is more stable under high temperatures, than its aluminous counterpart (Surdam 1973, Kuniyoshi & Liou 1976); although variables such as, high fO_2 (Liou 1979, Evarts & Schiffman 1983) and precursor mineral chemistry may be also be influential in controlling iron content. Iron-enriched prehnite within the Fishguard metabasites is restricted in general to cataclastites, suggesting the influence of high fO_2 and a_{Fe3+} within the fluid phase associated with tectonism.

The upper stability limit of prehnite has been experimentally determined at T=380°C at 2kbars P_{fluid} (Liou 1971). However, temperatures associated with the breakdown of prehnite, observed in 'transitional' assemblages in the Fishguard Volcanic Complex are likely to have been significantly less, as reaction temperatures are lower than the thermal stability limit of the reacting phase (Liou *et al.* 1985).

Plate 6.5 a + b. *FG 22 Prehnite cathode luminescense.* Comparative photomicrographs of coarse radiate prehnite from a basaltic cataclastite on the Pen Caer headland. Plate (**a**) standard transmitted light. Plate (**b**) the same crystal under cathode luminescence at an opereating voltage of 12-13kv. Such a technique highlights zoned crystal growth not apparent in standard microscopy, whilst variation in intensity of luminescence may relate to variation in chemistry. Preliminary analysis indicate the core (dark) is enriched with FeO* (assummed to be Fe^{3+}) to 4.3 wt% oxide, the top right hand margin of the crystal (light) contains FeO* at below detectable limits using energy dispersive spectroscopy. Whilst requiring further study, supression of luminesence in calcite is commonly attributed to iron-enrichment. Quartz occupies the non-luminescing background. Field of view 1.7mm.

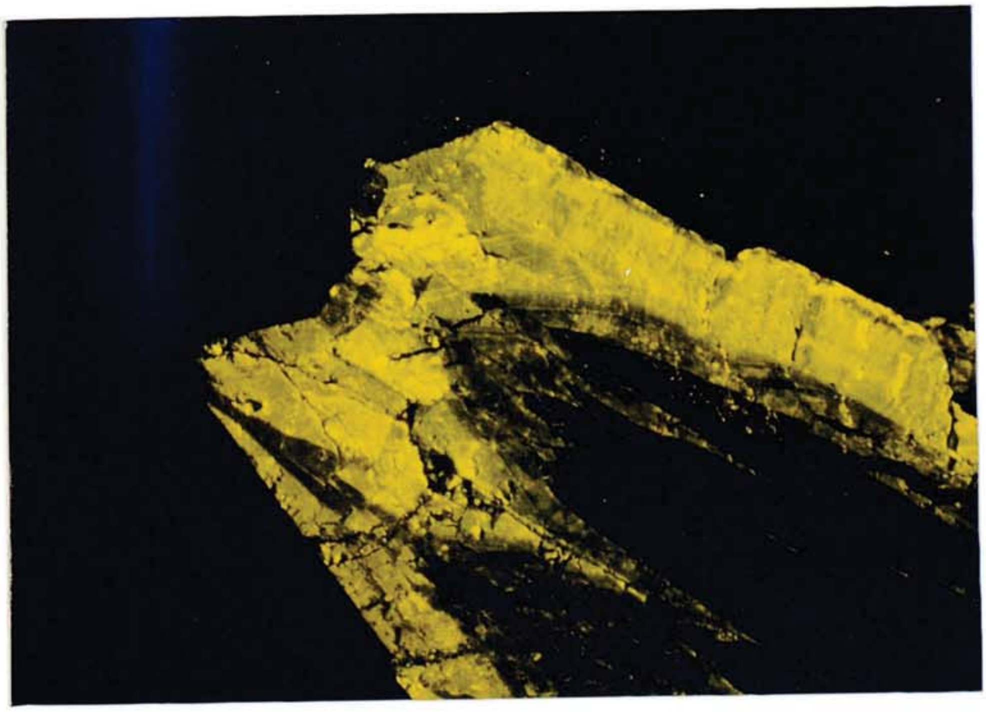
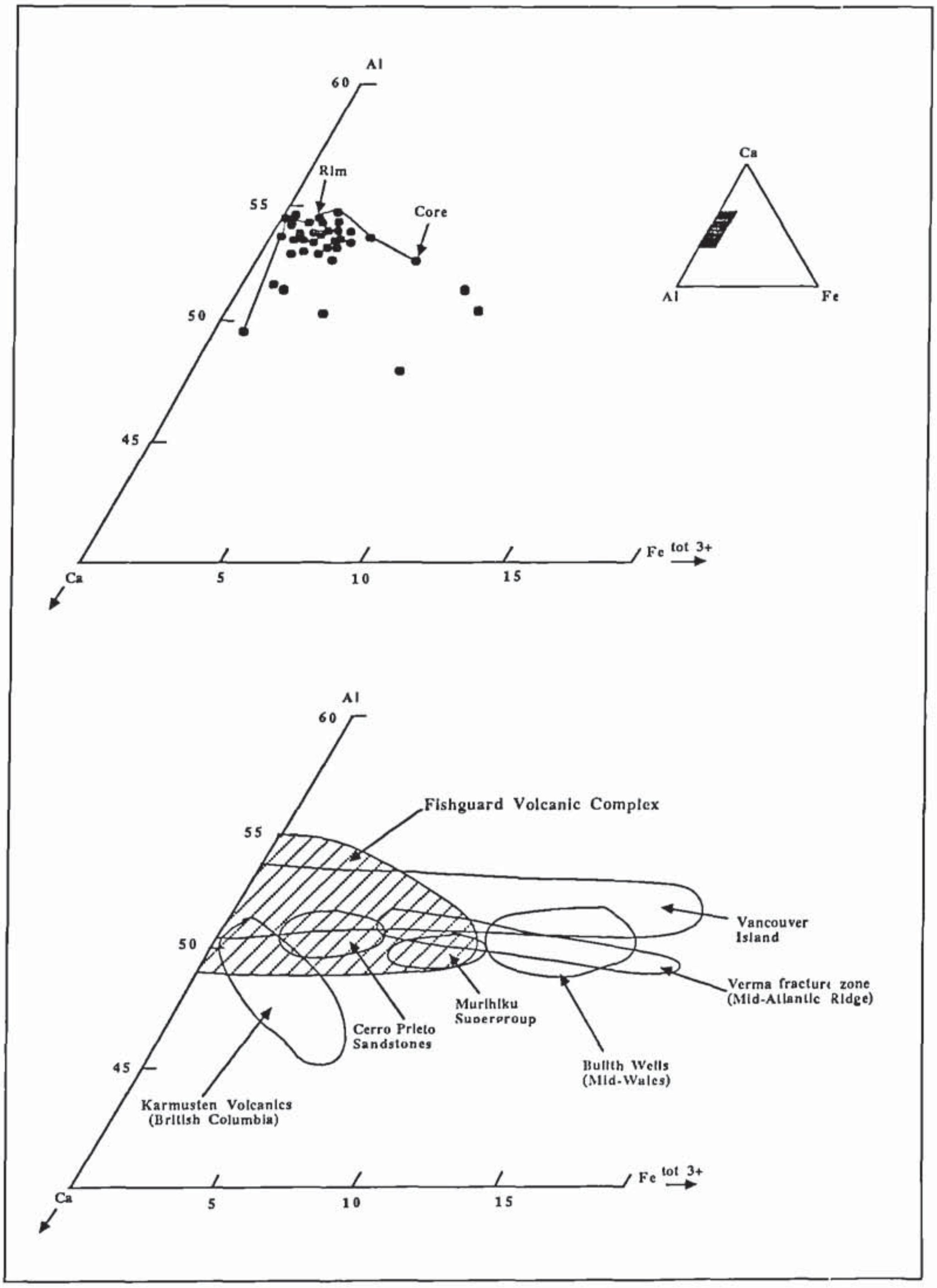


Figure 6.6. *Prehnite compositional diagram:* Triangluar Ca-Al-Fe^{3+} compositional diagrams of prehnite from the Fishguard metabastites. Fields in comparative diagram taken from (Karmusten Volcanics – Surdam 1969, Kuniyoshi & Liou 1976; Builth Wells – Bevins 1985; Verma Fracture Zone – Mevel 1980; Cerrio Prieto sndstonnes – Schiffman *et al.* 1985, Murihiku Super Group – Kay 1984)



Al
60
55
50
45
Rim
Core
Ca
5
10
15
Fe tot 3+
Ca
Al
Fe
Al
60
55
50
45
Fishguard Volcanic Complex
Vancouver Island
Verma fracture zone (Mid-Atlantic Ridge)
Builth Wells (Mid-Wales)
Murihiku Supergroup
Cerro Prieto Sandstones
Karmusten Volcanics (British Columbia)
Ca
5
10
15
Fe tot 3+

(e) Amphiboles.

Compositional, petrographic and textural relationships provide a good basis for distinguishing between the various calcic and sodic-calcic amphiboles observed during this study; although only one is here ascribed to development during low-grade metamorphism.

In an attempt to avoid the often cumbersome terminology of amphiboles, individual amphibole phases are discussed in terms of their dominant chemistry. The amphiboles identified, in probable paragenetic order are;

1) Kaersutite	**Magmatic**
2) Hornblende 3) Winchite (Sodic-calcic amphiboles) 4) Actinolitic hornblende	**Pre-regional alteration**
5) Actinolite	**Low-grade metamorphism**

Representative analyses of amphiboles are presented in Tables (6.8) and (6.9). Recalculation is based on 23 oxygens, according to the standard formula; $A_{0-1}B_2C_5{}^{vi}T_8{}^{iv}O_{22}(OH)_2$, where T= Si, Al^{iv} then Fe^{3+} and Ti^{4+}, C= Al^{vi}, Ti, Fe^{3+}, then Mg, Fe^{2+} and Mn, B= Fe^{2+}, Mn, Mg, then Ca and Na, A= Na and K (*c.f.* Leake 1978). Amphibole types 1, 2, 4, 5, (see above), conform to the constraints specified by Leake (1978) for calcic amphiboles ($(Ca+Na)_B \geq 1.34$ and $Na_B < 0.67$), whilst amphibole 3 conforms to the constraints specified for sodic-calcic amphiboles ($(Ca+Na)_B \geq 1.34$ and $0.67 \leq Na_B < 1.34$). Fe^{2+}/Fe^{3+} partitioning where appropriate has been carried out by adjusting the C and T cations to 13 (Leake 1978). This normalization effectively precludes Fe^{2+}, Mg and Mn from entering the B sites, although is thought to be appropriate as sufficient Ca and Na is invariably present to fill this site. Normalising calculations have not been carried out on kaersutite analyses due to the uncertainty of iron partitioning in this amphibole species (Rock & Leake 1984), although an assessment of the likely Fe^{3+} content is made (see below). For convenience, amphiboles are discussed in likely paragenetic order, the peripheral nature of kaersutite and deuteric/hydrothermal amphiboles precludes detailed discussion.

(e).1. Kaersutite

Kaersutite has been observed as small (0.2 - 1mm) subhedral to anhedral poikilitic inclusions within clinopyroxenes from gabbros around Llwanda and Garn Fawr. Its occurrence is sporadic, although locally it may be abundant. It may exhibit epitaxial actinolite overgrowths, and/or, be replaced by chlorite.

Analyses (Table. 6.7.) reveal that they classify as kaersutite (Fig. 6.7.a.) and accord to the parameters specified by Leake (1978), where; $(Ca + Na)_B \geq 1.34$, $Na_B < 0.67$ and $Ti \geq 0.50$).

Table 6.7 Representative Amphiboles from the Fishguard District (Kaersutite, Hornblende, Actinolitic Hornblende, Sodic Calcic Amphibole)

Class	K	K	K	K	FFAHC	FFAH	FA	FFAH	FAHC	FFA	FFAH	FFAH	FoW	FerW	Ric	Win
SiO_2	40.16	40.04	39.39	40.71	50.58	49.82	50.70	49.73	49.82	51.12	50.00	50.98	52.86	54.59	54.56	54.76
TiO_2	6.09	7.81	8.43	6.93	0.27	0.25	n.d.	0.27	n.d.	n.d.	n.d.	n.d.	0.57	0.64	0.65	0.42
Al_2O_3	14.49	12.61	12.95	12.70	2.53	3.00	2.36	3.05	2.76	2.60	2.60	2.63	0.79	n.d.	n.d.	0.39
FeO^*	12.56	15.69	16.02	14.61	27.49	26.56	23.46	24.97	25.97	24.68	24.41	24.45	24.32	21.37	19.85	19.20
MnO	n.d.	n.d.	n.d.	n.d.	0.39	0.51	0.54	0.48	0.68	0.66	0.62	0.46	0.77	0.85	0.51	0.64
MgO	11.33	9.80	9.29	10.21	7.15	7.13	9.44	8.27	7.98	8.64	8.75	9.06	7.76	10.25	10.86	11.23
CaO	11.35	11.50	12.00	11.66	8.74	9.89	10.10	9.66	8.20	9.91	10.30	10.55	6.86	4.98	5.12	5.27
Na_2O	2.68	2.31	2.48	2.50	1.70	1.35	1.32	1.46	1.82	1.19	1.14	1.50	3.37	4.34	5.73	4.57
K_2O	0.11	0.18	0.12	0.16	n.d.	n.d.	0.15	0.13	n.d.	0.17	n.d.	n.d.	0.69	1.07	1.08	0.64
Total	98.77	99.94	100.68	99.48	98.85	98.51	98.07	98.02	97.23	98.97	97.82	99.63	97.99	98.09	98.36	97.12
Recalculation based on 23 Oxyens																
Si	5.870	5.883	5.772	5.973	7.494	7.437	7.521	7.419	7.422	7.530	7.476	7.495	7.919	7.979	8.035	7.990
Al^{iv}	2.130	2.117	2.228	2.127	0.442	0.528	0.413	0.537	0.485	0.452	0.458	0.456	0.081	0.000	0.000	0.010
Fe^{3+}	n.c.	n.c.	n.c.	n.c.	0.064	0.035	0.066	0.044	0.093	0.018	0.066	0.049	0.000	0.021	0.000	0.000
ΣT	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.035	8.000
Al^{vi}	0.367	0.067	0.009	0.069	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.059	0.000	0.000	0.090
Ti	0.670	0.863	0.930	0.765	0.030	0.028	0.000	0.030	0.000	0.000	0.000	0.000	0.064	0.070	0.072	0.046
Fe^{3+}	n.c.	n.c.	n.c.	n.c.	1.182	0.952	0.860	0.986	1.434	0.972	0.915	0.753	0.581	0.912	0.330	0.698
Fe^{2+}	1.494	1.923	1.963	1.793	2.159	2.328	1.984	2.083	1.707	2.049	2.066	2.203	2.465	1.699	2.113	1.644
Mg	2.469	2.147	2.029	2.234	1.580	1.627	2.089	1.840	1.773	1.898	1.949	1.986	1.733	2.234	2.385	2.444
Mn	0.000	0.000	0.000	0.000	0.049	0.065	0.067	0.061	0.086	0.081	0.078	0.057	0.098	0.105	0.064	0.079
ΣC	5.000	5.000	4.931	4.861	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	4.964	5.001
Fe^{2+}	0.041	0.005	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Ca	1.779	1.811	1.884	1.833	1.387	1.582	1.605	1.544	1.309	1.564	1.648	1.662	1.101	0.780	0.808	0.824
Na	0.180	0.474	0.116	0.167	0.488	0.380	0.380	0.422	0.526	0.340	0.330	0.338	0.899	1.220	1.192	1.176
ΣB	2.000	2.000	2.000	2.000	1.876	1.972	1.985	1.966	1.834	1.904	1.978	2.000	2.000	2.000	2.000	2.000
Na	0.581	0.474	0.590	0.544	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.089	0.089	0.009	0.444	0.156
K	0.022	0.035	0.024	0.031	0.000	0.000	0.028	0.025	0.000	0.032	0.000	0.000	0.123	0.200	0.203	0.151
ΣA	0.603	0.509	0.614	0.575	0.000	0.000	0.028	0.025	0.000	0.032	0.000	0.089	0.212	0.209	0.647	0.307

Class	FoR	FeW	FoW	FeW	FoW	FoRi	FerFoW	FerFoW	FrScMH	FeMH	FenMH	FeMH	FiFoH	FiScMH	FiScMH	
SiO_2	53.48	54.04	55.19	51.13	50.32	52.76	51.28	52.83	48.57	48.51	48.91	47.38	47.17	48.16	47.28	47.92
TiO_2	1.26	0.00	0.84	1.51	1.87	0.00	1.08	1.16	1.11	1.34	1.12	1.26	0.77	1.10	1.46	1.39
Al_2O_3	0.39	0.67	n.d.	1.13	1.58	1.12	1.01	0.98	3.41	4.36	4.46	4.13	3.37	3.24	3.61	3.02
FeO^*	20.54	20.39	18.70	26.33	26.82	24.73	26.24	23.53	22.63	18.55	20.02	20.50	28.33	23.82	24.07	23.16
MgO	10.02	10.81	12.51	6.52	5.95	9.15	7.29	8.43	10.36	13.13	11.55	11.96	6.22	10.10	9.75	10.11
MnO	n.d.	0.69	n.d.	0.81	0.57	0.58	0.77	0.87	0.42	0.36	0.51	0.57	0.43	0.47	0.39	0.42
CaO	6.05	7.12	6.70	5.01	4.97	7.16	5.58	5.11	9.44	9.80	10.66	10.26	10.06	9.40	9.16	9.25
Na2O	4.54	3.34	3.89	4.64	4.68	3.10	4.76	4.61	2.21	2.45	1.66	2.39	1.11	2.22	2.76	2.72
K_2O	1.19	0.63	0.83	1.51	1.13	0.25	0.97	1.01	0.67	0.37	0.64	0.65	0.40	0.63	0.84	0.63
Total	97.50	97.39	98.66	98.15	97.89	98.85	98.98	98.53	98.82	98.96	99.53	99.10	97.86	99.14	99.32	98.62
Recalculation based on 23 Oxygens.																
Si	7.991	7.931	7.962	7.727	7.663	7.667	7.685	7.819	7.163	7.003	7.112	6.948	7.210	7.100	7.009	7.134
Al^{iv}	0.069	0.116	0.000	0.201	0.284	0.219	0.182	0.179	0.593	0.742	0.765	0.714	0.608	0.536	0.663	0.530
Fe^{3+}	0.000	0.000	0.038	0.072	0.053	0.114	0.133	0.002	0.244	0.255	0.123	0.338	0.182	0.346	0.328	0.336
ΣT	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al^{vi}	0.060	0.047	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Ti	0.142	0.000	0.091	0.104	0.241	0.228	0.122	0.129	0.123	0.146	0.122	0.139	0.089	0.122	0.162	0.156
Fe^{3+}	0.178	0.771	0.543	0.740	0.686	0.221	0.714	0.797	0.850	0.895	0.736	0.750	0.911	0.924	0.821	0.687
Fe^{2+}	2.388	1.730	1.674	2.514	2.675	3.106	2.440	2.112	1.696	1.088	1.575	1.425	2.525	1.665	1.821	1.859
Mg	2.834	2.366	2.692	1.470	1.351	1.361	1.629	1.861	2.279	2.827	2.505	2.615	1.418	2.221	2.146	2.245
Mn	0.000	0.086	0.000	0.104	0.074	0.085	0.098	0.109	0.053	0.044	0.063	0.071	0.056	0.059	0.049	0.053
ΣC	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000	5.000
Fe^{2+}	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Ca	0.968	1.119	1.036	0.811	0.811	0.981	0.896	0.810	1.491	1.516	1.661	1.612	1.647	1.485	1.449	1.475
Na	1.032	0.881	0.964	1.189	1.189	1.109	1.104	1.190	0.509	0.484	0.339	0.388	0.329	0.515	0.551	0.525
ΣB	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	2.000	1.976	2.000	2.000	2.000
Na	0.292	0.070	0.123	0.205	0.192	0.456	0.279	0.129	0.123	0.227	0.128	0.291	0.000	0.119	0.239	0.272
K	0.227	0.062	0.153	0.172	0.220	0.221	0.185	0.191	0.126	0.068	0.119	0.122	0.078	0.119	0.158	0.120
ΣA	0.519	0.132	0.276	0.377	0.412	0.677	0.464	0.320	0.249	0.295	0.247	0.413	0.078	0.238	0.397	0.392

Table 6.7. Representative analysis of pre-regional amphiboles from the Fishguard metabasites. (Brown Hornblende, Blue-green actinolitic Hornblende, Kaersutite, Winchite/Ricterite)

The over-riding factor in distinguishing kaersutite, being Ti ≥ 0.50 (Rock & Leake 1984). Analyses presented in Table (6.7.) are from individual crystals within one thin-section, indicating significant intergranular variation with regards Ti, Al^{vi}, and Fe* - Mg. Due to the fine grain size, zonation if present could not be detected. Analyses are in general comparable to those recorded in Deer *et al.* (1962), although total iron is slightly higher.

In addition to simple Fe^{2+}<—> Mg exchange coupled substitutions are apparent, the more important being; edenite (Na, K(A) + Al (T) for ˇ A + Si (T)), tschermarkite (Al, Fe^{3+} (C) + Al (T) for Mg, Fe (C) + Si (T)), and Ti-tschermarkite (Ti (C) + 2 Al (T) for Mg, Fe (C) + 2 Si (T)). Robinson *et al.* (1971) noted that a combination of the above (i.e. A-site + Al (C) + Fe^{3+} (C) + 2 Ti *vs.* Al (T)) should yield a 1:1 slope. By using this combination, it is possible to evaluate the Fe^{3+} content of the analysed kaersutite by assuming Fe^{3+} to be zero (i.e. A site + Al (C) + 2Ti *vs.* Al (T)), even though uncertainty exists as to the iron oxidation partitioning in this amphibole species (Rock & Leake 1984). Using this assumption (i.e. Fe^{3+}= 0), analysed kaersutites from the Fishguard volcanics, lie on or lie close to, the 1:1 slope of the plot A-site + Al (C) +2Ti *vs.* Al (T) (Fig. 6.7.c.), suggesting that whilst the Fe^{2+}:Fe^{3+} ratio has not been calculated, Fe^{3+} is likely to be absent or present in minor amounts.

Kaersutite is a common minor constituent of a variety of rock types, such as; camptonites (Scott 1980), alkali wherlites (Lippard 1984) and layered basic intrusives (Brown *et al.* 1982). Titanium–rich amphiboles are commonly regarded as of magmatic affinity, this being the most probable explanation for the poikilitic kaersutites observed, possibly crystallising as an early magmatic phase prior to the development of clinopyroxene.

(e).2. Brown Hornblende

Brown hornblende is widely developed throughout the gabbroic rock of the Pen Caer peninsular. However, its occurrence in thin-section is sporadic, attributable in part to replacement by other phases. Hornblende most characteristically occurs as anhedral plates in interstitial chlorite, although is observed locally as subhedral rhombs. It may partially replace clinopyroxene and more rarely ore, although not feldspar. In all observed occurrences it shows evidence for at least partial 'uralic' replacement by green and dark-green actinolitic amphibole, variable from slight alteration along terminal edges to near total replacement, where only remnent variegated patches of hornblende are preserved. This coexistence is not thought to represent a miscibility gap (*c.f.* Klien 1969), the existence of which is a matter for debate (e.g. Grapes & Graham 1978, Hayes 1982). Rather it is thought, that coexistence represents phase development during two separate events of alteration. Evidence supporting two events is based on textural observation; a) contacts between actinolite and hornblende are diffuse and variegated, b) no separate homogenous grains or fibres are observed in contact, c) no exsolution lamellae of actinolite in hornblende, or visa versa, are seen (this is however a rare phenomenon with only a few examples documented e.g. Hall 1985), d) a general lack of optical continuity, e) kaersutite and sodic-calcic amphiboles are replaced in a similar manner; a coexistence which is difficult to interpret in terms of a solvus. The fine grain size of actinolitic amphibole replacing hornblende has precluded a detailed study of the replacing phase, and must await further study using coarser grained material.

Representative chemical analyses of hornblende are presented in Table (6.7.). Assuming Fe^{2+}:Fe^{3+} partitioning is correct, most classify as magnesio-hornblende and a few as ferro-hornblende (Fig. 6.7.a.). According to the classification of Leake (1978), most earn the additional prefix of ferri ($0.75 > Fe^{3+} < 0.99$) or ferrian ($Fe^{3+} > 1.00$), serving as an indication as to the unusually high Fe^{3+} content. Whilst the Fe^{3+} content can not be categorically proven, the occurrence of Al only in the T-site suggests that Fe^{3+} is a dominant R^{3+} cation in the C-site. Si and Al^{vi} occupancy of the T-site is invariably less than 8, requiring in theory Fe^{3+} to enter this site; although noting the reservation of Hawthorn (1981), who points out that there is no convincing evidence for tetrahedrally co-ordinated Fe^{3+} in amphiboles. Total iron is variable, although characteristically high (19.7 - 28.3 wt. % oxide), with Mg in part sympathetic accountable to simple Fe^{2+}<—> Mg exchange. Na_A is variable and commonly approaches the classification parameters of edenitic amphiboles (i.e. A-site > 0.5), with K accounting for up to 0.175 of the A-site cations. TiO_2 is variable (0.52 to 1.84 wt. % oxide), although this is consistent with secondary hornblende from amphibolitised metagabbros subjected to ocean-floor metamorphism (e.g. Cortesogno & Lucchetti 1984, Sievell 1984). Intergranular variation is on the whole limited, whilst intra- and interspecimen variation may be considerable. In addition to simple Fe^{2+} <—> Mg exchange, coupled substitutions are apparent, the more important being, ferri-tschermakite (Fe^{3+} (C) + Al (T) for Si (T) + Mg (C)), riebeckite (Na (B) + Fe^{3+} (C) for Ca (B) + Mg (C)) and edenite (Na, K (A) + Al (T) for ˇ A + Si (T)). Whilst ambiguity exists as to tetrahedrally co-ordinated Fe^{3+}, it would appear that a ferri-ferri tschermarkite substitution common to pyroxenes may also be appropriate.

Many studies have stressed the influence of metamorphic conditions on amphibole chemistry, particularly with regards Al^{iv}/Al^{vi} partitioning, Na_A, Si, and Ti content (e.g. Leake 1965, Raase 1974, Laird & Albee 1981). However, based on comparisons of published amphibole data from low and medium pressure facies, Hynes (1982) suggests that a systematic variation in Na, Na_A, Al or Al^{vi}, can not be recognised as a function of pressure, although he does believe that a distinction between low and medium pressure facies series can be made by variation in the Ti/Al ratio of amphiboles. The Ti/Al ratio of analysed hornblendes (Table 6.7.), all plot in the low-pressure facies field of Hynes's (1982) Ti/Al discriminatory diagram, which suggests that crystallisation occurred at low-pressure. This is supported by using the Al^{vi} +Fe^{3+}

+2Ti *vs.* Na_{M4} plot of Laird *et al.* (1984), with all amphiboles ploting in the low pressure facies series field (Fig. 6.7.d.). Low pressures would also be consistant with low Al^{vi} and high Na_A (Leake 1965, Laird & Albee 1981) and stratigraphic relationships, of a post-magmatic phase developing in high level gabbroic intrusions.

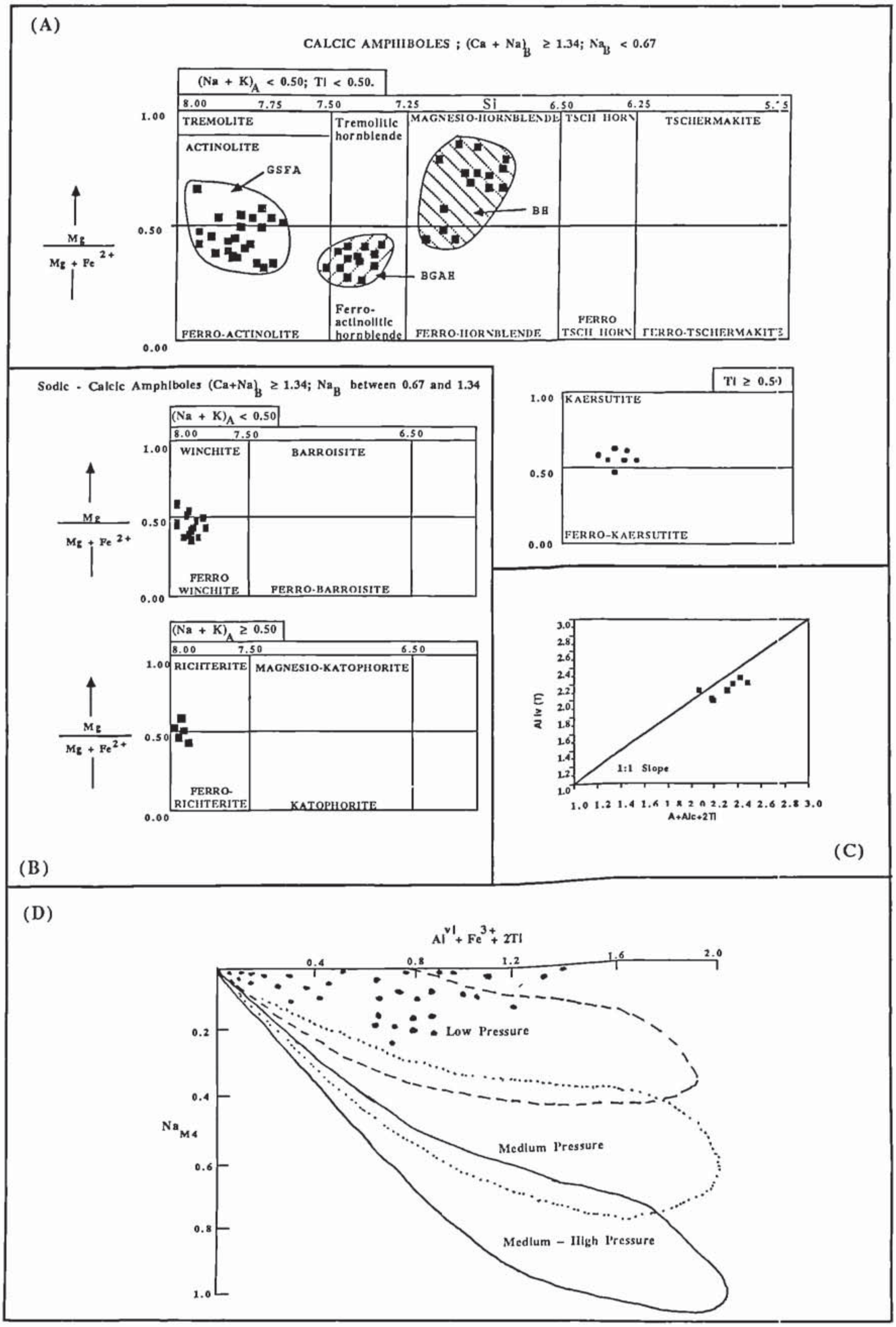
Whilst hornblende crystallised as a post-magmatic phase prior to low-grade regional metamorphism, it is uncertain as to whether it represents the product of deuteric fluids or as the result of hydrothermal circulation. In several instances there is evidence for mutual coexistence with quartz containing occasional apatite rhombs, possibly indicating alteration by late stage magmatic fluids enriched in volatiles. However, analyses of hornblende indicate unusually high Fe^{3+}, Mg, and Si, cations favoured by high fO_2 (Spear 1981), suggesting an oxygen enriched fluid and high a_{Fe3+} at the time of crystallisation. Whilst speculative, given the high–level nature of the gabbroic intrusions beneath a syn-generic volcanic pile, hornblende may have crystallised as a result of permeating seawater passing from surfacial levels downwards through permeable strata; analogous to submarine hydrothermal amphibolitisation of high level oceanic gabbros.

(e).3&4. Lilac sodic-calcic amphibole & Blue-green actinolitic hornblende.

A further suite of secondary amphiboles are developed in association with felsic microdykes cross-cutting the gabbros in the vicinity of Llwanda and Carn Gelli {5.4.2.}, occurring within the microdyke groundmass and within the peripheral margins of the host gabbro. Optically and chemically two amphibole 'species' are identified; lilac sodic-calcic amphiboles of the winchite-richterite series (and Fe equivalent), and blue-green actinolitic hornblende; the latter is however, not exclusive to the gabbros of Llwanda and Carn Gelli.

Blue-green actinolitic Hornblende; Blue-green actinolitic hornblende, occurs as mat-like acicular bundles within the groundmass, and more rarely as epitaxial overgrowths on clinopyroxene, its form having similar characteristics to that of actinolite (see below). It is however readily identifiable from actinolite by its deep blue-green colouration, strong pleiochroism and anomalous interference colours. Analysis of blue-green actinolitic hornblende are presented in Table (6.7.). As the name implies compositions are intermediate between actinolite and hornblende, analyses plotting in the ferro-actinolitic hornblende field (Fig. 6.7.b.). The general chemical characteristics of the blue-green actinolitic hornblende are that most earn the additional prefix of ferri ($0.75 > Fe^{3+} < 0.99$) or ferrian ($Fe^{3+} > 1.00$), assuming Fe^{2+}/Fe^{3+} partitioning is correct. Total iron is extremely high, in the range 23.46 to 27.49 wt. % oxide. Si and Al^{vi} in the T-site are generally less than 8, requiring small quantities of Fe^{3+} to enter this site. Na_B is variable from 0.33 to 0.53, whilst the A-site is generally vacant. TiO_2 is

Figure 6.7. *Amphibole compositional diagrams.* **(A).** Si *vs.* $Mg/Mg+Fe^{2+}$ plot for Calcic amphiboles ($((Ca+Na)_B \geq 1.34$; $Na_B < 0.67$; Kaersutite where $Ti \geq 0.5$). Fields show are for greenschist facies actinolites (GSFA); Blue-green actinolitic Hornblende (BGAH) and Brown Hornblende (BH). **(B).** Si *vs.* $Mg/Mg+Fe^{2+}$ for sodic-calcic amphiboles ($((Ca+Na)B \geq 1.34$; Na_B between 0.67 and 1.34.) from felsic sills cross-cutting gabbros at Henner Cross [S.M.]. **(C).** $Al^{iv}(T)$ *vs.* A-site +Al (C) + 2Ti for analysed kaersutites. **(D).** Discrimonatory plot of Al^{vi} + Fe^{3+} + 2Ti *vs.* Na_{M4} for all amphiboles within the Fishguard district.



(A)
CALCIC AMPHIBOLES ; (Ca + Na)B ≥ 1.34; NaB < 0.67
(Na + K)A < 0.50; Ti < 0.50.
Si
TREMOLITE
ACTINOLITE
Tremolitic hornblende
MAGNESIO-HORNBLENDE
TSCH HORN
TSCHERMAKITE
GSFA
BH
BGAH
Ferro-actinolitic hornblende
FERRO-ACTINOLITE
FERRO-HORNBLENDE
FERRO TSCH HORN
FERRO-TSCHERMAKITE
Mg / Mg + Fe 2+
Sodic - Calcic Amphiboles (Ca+Na)B ≥ 1.34; NaB between 0.67 and 1.34
(Na + K)A < 0.50
WINCHITE
BARROISITE
FERRO WINCHITE
FERRO-BARROISITE
(Na + K)A ≥ 0.50
RICHTERITE
MAGNESIO-KATOPHORITE
FERRO-RICHTERITE
KATOPHORITE
(B)
Ti ≥ 0.50
KAERSUTITE
FERRO-KAERSUTITE
Al iv (T)
1:1 Slope
A+Alc+2Ti
(C)
(D)
Al vI + Fe 3+ + 2Ti
NaM4
Low Pressure
Medium Pressure
Medium - High Pressure

generally absent, although is recorded in trace amounts up to 0.49 wt.% oxide.

Lilac sodic-calcic amphiboles; Whilst blue-green actinolitic hornblende is identified outside of felsic micro-dykes, lilac sodic-calcic amphiboles of the winchite-richterite series appear to be wholly restricted to this lithology. Lilac sodic-calcic amphiboles occur as fine subhedral rhombs and variegated aggregates within the microdyke groundmass, and peripheral groundmass of the host gabbro, commonly corroded and replaced along terminal edges by actinolite. The colour of the amphibole is invariably lilac, with a moderate pleiochroic scheme (lilac to pale blue-lilac), colouration probably reflecting their unusually high iron content. Interference colours are generally anomalous orange-blue.

Analyses of lilac sodic-calcic amphiboles are presented in Table 6.7. All analyses plot within the sodic-calcic amphibole field of the winchite-richterite series (and Fe-equivalents, Fig. 6.7.b.). A broad chemical variation is observed from all the analysed crystals, although little inter-crystalline variation was noted, individual crystals appearing to be homogenous. Whilst there is no petrographic evidence for a mixed phase or apparent zonation, the fine grain size and frequent replacement by actinolite, has however, precluded a detailed microprobe traverse across single crystals; this must await further study. The general chemical characteristics of the lilac sodic-calcic amphiboles analysed are that Al^{iv} is required in small, although variable, amounts to fill the T-site, Si cation content being high. Total iron is unusually high, although variable from 18.70 - 26.82 wt. % oxide. Several analyses earn the additional prefix 'potassian', where K = 0.25 - 0.49, whilst the majority earn the additional prefix 'ferro' where Mg/(Mg+Fe) < 0.50.

(e).5. Actinolite

Whilst the distribution of 'pre-regional' amphiboles is generally restricted to the larger gabbroic bodies of the Pen Caer peninsular, actinolite is widely developed throughout the district in typical greenschist assemblages. Its common replacement of 'pre-regional' amphiboles, attests to its later origin, whilst its development independent of relic phases in prograde 'transitional' assemblages and greenschist facies assemblages, precludes a post-magmatic deuteric or hydrothermal alteration for the majority of occurances.

Actinolite is most characteristically observed as;

a) fine acicular needles and bi-pyramidal rhombs in a chloritic groundmass
b) fine epitaxial overgrowths on relict clinopyroxene (Uralitization),
c) a replacement of feldspar and clinopyroxene along structural weaknesses,
d) variegated patches along the terminal edges of pre-existing amphiboles
e) fine acicular aggregates radiating from highly corroded aggregates of prehnite.

In thin-section actinolite is variable in colour from light–green to colourless, a moderate pleiochroism may be observed in more strongly coloured varieties. A discussion regarding the chemistry of actinolite, is restricted to those analysed from typical greenschist or 'transitional' type assemblages, as the compositions of actinolites replacing pre-existing amphiboles appear to have inherited in some instances, small characteristics of the amphibole they replace. This is exemplified by actinolite replacing sodic-calcic amphiboles, where the Na content of the replacing actinolite is at variance with those from the low grade assemblages; and is not taken to represent an indication of medium to high pressures (Hynes 1982). The inheritance within actinolite of precursor amphibole chemistry is also suggested by the one reliable analysis of actinolite replacing kaersutite, were 0.35 wt.% TiO_2 was recorded.

The general characteristics of actinolite from greenschist and transitional assemblages, are typical of those documented from other low-grade metamorphic terrains (see Kay 1983 Coombs *et al.* 1976). Si in the T-site is always greater than 7.65 cations per formula unit, Al^{iv} being sufficient in all analyses to fill the T-site to a maximum of 8 (Fig. 6.7.a., Table 6.8.). Many analysed actinolites classify as ferro-actinolites, were $Fe^{2+}/Fe^{2+} + Mg < 0.50$ (Fig. 6.7.a.). Total Al_2O_3 varies from 0.32 to a maximum of 4.46 wt.% oxide, although is generally less than 2 wt.% oxide; low Al_2O_3 content is typical of actinolites developed in low grade terrains (e.g. Kuniyoshi & Liou 1976, Kay 1983). Al^{vi} is present in small, though variable amounts, possibly indicating minor tschermarkitic substitution, although the dominant substitution appears to be simple $Fe^{2+} \longleftrightarrow Mg$ exchange. A-site occupancy is generally absent, whilst the minor presence of Na_A may indicate a limited richterite substitution. TiO_2 is invariably absent, or present in trace amounts, the widespread presence of sphene suggests however, that the amphibole is always saturated with respect to Ti. Low TiO_2 contents being typical of actinolites from low grade terrains (Liou & Ernst 1979, Hynes 1982).

Whilst being readily observed, the fine grain size of many actinolites has precluded a detailed study, coarse grained aggregates being rare. Analysed actinolites are comparable to those recorded from greenschist facies assemblages elsewhere in the Welsh Basin (see Bevins & Rowbotham 1983), and compare favorably with other low grade terrains.

(f). Epidote

Epidote group minerals are common to many rock types within the Fishguard Volcanic Complex. Epidote (piscasitic Fe-rich) is extensively developed in metabasites, whilst clinozoisite has been observed within sedimentary hosted cataclastites. Zoisite has not been observed. Epidote can be observed in a variety of forms, although most characteristically occurs as;

a) euhedral rhombs within areas of interstitial mesostasis (Plate 6.1.)

b) granoblastic and radiate aggregates in basic tuffs and hyaloclastites (Plate 6.6.)
c) a vein-filling mineral and constituent of vesicular assemblages in pillow lavas
d) amorphous cryptocrystalline aggregates within interstitial chlorite

Epidote group minerals have the general formula $X_2Y_3Si_3O_{12}$ (OH), defined by the binary components, zoisite ($Ca_2Al_3Si_3O_{12}$(OH)) and pistacite ($Ca_2(Fe^{3+})_3Si_3O_{12}$ (OH)). The strong tendency for Fe^{3+} to be partitioned into only one of the three distinct octahedral sites, invariably leads to naturally occurring epidotes exhibiting no more than 33 mol% solid solution of $Ca_2Fe^{3+}{}_3Si_3O_{12}$ (OH) in $Ca_2Al_3Si_3O_{12}$ (OH). This phenomenon is adhered to in the majority of analysed epidotes from the Fishguard metabasites (Figure 6.8.b.), although unusually Fe-rich epidotes (≈ 40 mol% Ps) are recorded from a cataclastite hosted within pillow lavas at Carn Hendi [S.M. 9370 3885]. Epidote analyses from metabasites, metabasite cataclastites, and clinozoisite from a sedimentary cataclastites (clinozoisite 0-15% Ps in the monoclinic structure, Dollase 1968) are presented in Table (6.9.). Recalculation is based on 12.5 oxygens and all iron as Fe_2O_3, following the general formula X_2Y_3 (Si (+Al)$_3O_{12}$ (OH)); where, X= Ca and Mg, Y= Fe^{3+}, Al, and Mn.

The general character of metabasite epidotes are; that tetrahedrally co-ordinated Al is absent as Si is always in slight excess of 3 (Table 6.9.), the dominant octahedral substitution is Fe^{3+}<—> Al in the Y-site; Ca is generally in slight excess of 2, although minor Mg<—>Ca substitution is apparent to several analyses. The pistacite component of epidotes from metabasites is limited in the range 29.0 – 35.4 mol % (Ps), the majority being less than or close to the theoretical $Ca_2(Al_2,Fe^{3+})Si_3O_{12}$ (OH). There appears to be no direct variation in Ps% component within epidotes from prehnite-pumpellyite, 'transitional', or greenschist assemblages. Intergranular variation is limited to a maximum of 3.5 mol % (Ps). Complex zonation commonly observed in epidotes from low-grade terrains (e.g. Nakajima *et al.* 1977, Cho *et al.* 1986) appears to be absent outside of metabasite cataclastites (see below).

Bevins & Rowbotham (1983), record zoned epidotes from dolerites in the Prescelly Hills, containing Fe-rich cores and Al-rich rims, which they attribute to crystallisation under prograde conditions; such zonation has been used by many as supportive evidence for delimiting prograde conditions (e.g. Brown 1967, Kawachi 1975, Smith *et al.* 1982). Cooper (1972) proposed that Fe-rich epidotes typically develop under high fO_2, whilst Coombs *et al.* (1976) suggest that Fe-depletion frequently observed in zonation may be accountable to the general tendency of decreasing fO_2 and additional availability of $Ca_2Al_3Si_3O_{12}$(OH) with advancing metamorphism. The high iron content of epidotes recorded here, with no obvious disparity between varying assemblages, may therefore indicate that fO_2 remained high and relatively constant under progressive conditions. This may suggest that syn-metamorphic tectonism and high permeability allowed the widespread circulation of fluid, keeping oxygen fugacity high

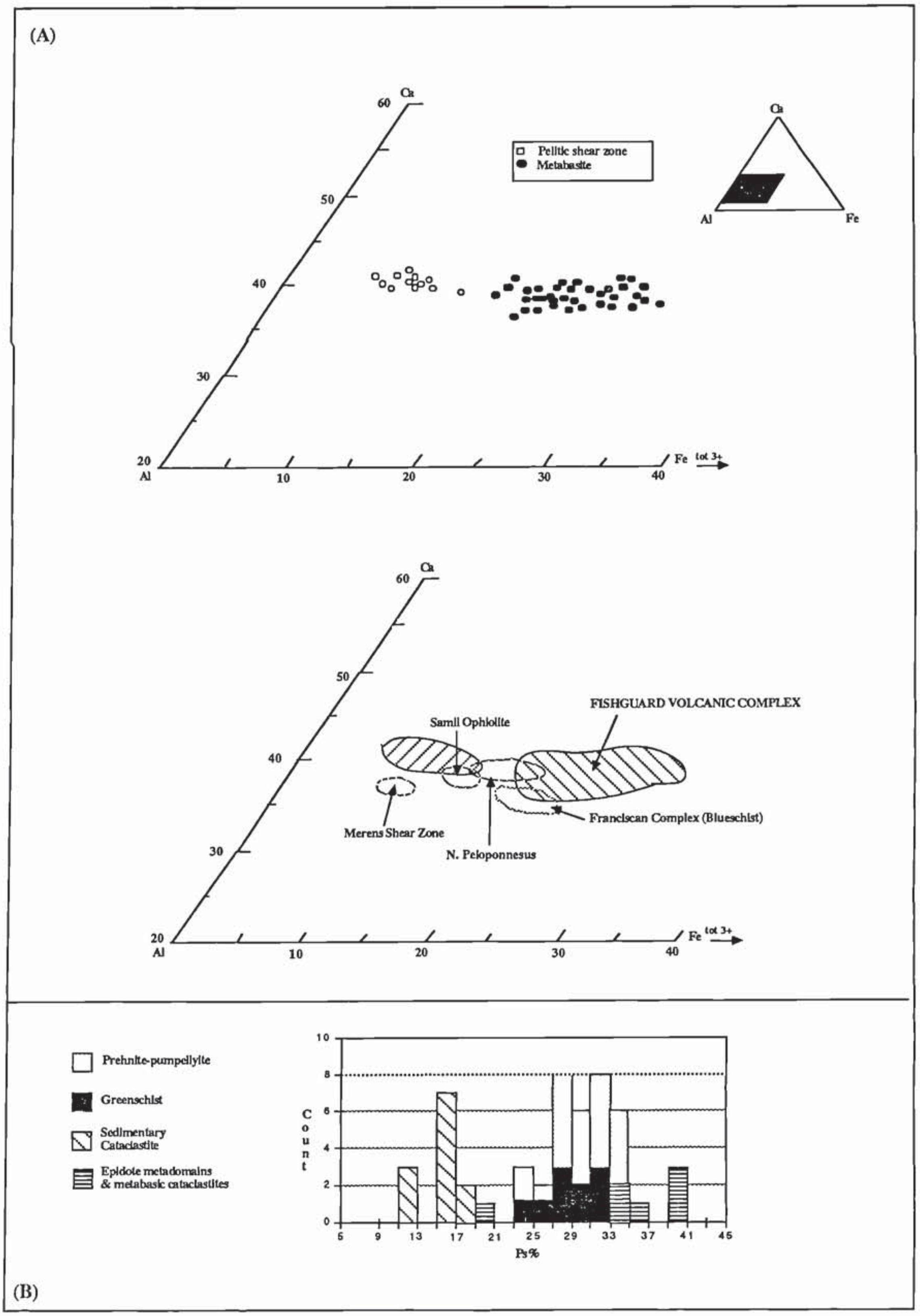
Table 6.8. Actinolite analyses from the Fishguard District

SiO_2	52.76	53.53	53.36	52.67	53.17	53.67	53.11	52.44	53.75	52.85	53.85	52.07	54.98	53.61	53.54	52.76	54.69	53.20
TiO_2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	0.35	n.d.	n.d.	n.d.
Al_2O_3	1.56	2.30	0.93	1.70	1.68	1.68	2.06	1.72	1.58	1.29	0.32	1.55	4.46	4.61	2.85	4.15	0.84	1.15
FeO*	21.09	19 85	20.02	19.78	20.85	17.69	16.79	16.18	16.38	20.30	16.67	19.88	15.05	16.05	16.49	18.15	14.21	18.03
MnO	0.31	0.29	0.41	0.62	n.d.	0.47	0.21	0.21	0.72	0.34	n.d.	0.33	n.d.	0.24	0.30	n.d.	0.41	0.38
MgO	10.32	9.64	10.69	10.76	10.18	12.42	13.26	13.24	13.67	10.10	12.62	11.97	10.45	10.51	12.52	11.37	15.06	12.44
CaO	12.42	12.00	12.61	12.56	12.57	12.72	12.43	12.46	12.61	11.76	12.78	12.30	9.98	11.46	11.97	9.52	12.66	12.53
Na_2O	0.74	1.15	n.d.	n.d.	0.34	0.33	0.24	0.33	0.40	0.76	n.d.	n.d.	2.25	1.87	n.d.	1.06	0.46	0.41
K_2O	n.d.	n.d.	0.16	0.12	n.d.	0.12	0.16	0.06	0.14	0.72	n.d.	n.d.	0.28	0.12	n.d.	0.55	n.d.	0.17
Total	99.20	98.76	98.18	98.21	98.79	99.10	98.26	96.64	99.25	98.12	96.24	98.10	97.45	98.47	98.02	97.56	98.33	98.31
Structural formulae on the basis of 23 O																		
Si	7.799	7.867	7.918	7.801	7.853	7.808	7.709	7.781	7.702	7.910	8.006	7.636	7.950	7.865	7.618	7.638	7.868	7.815
Al^{iv}	0.201	0.124	0.082	0.199	0.147	0.192	0.291	0.219	0.267	0.090	0.000	0.268	0.050	0.135	0.382	0.342	0.132	0.185
ΣT	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	7.969	8.000	8.006	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al^{vi}	0.071	0.276	0.081	0.089	0.146	0.096	0.062	0.079	0.000	0.138	0.056	0.000	0.710	0.614	0.096	0.348	0.100	0.140
Ti	0.000	0.000	0.000	0.000	0.000	0.000	0.004	0.004	0.000	0.000	0.000	0.000	0.000	0.000	0.037	0.000	0.000	0.000
Fe^{3+}	0.000	0.000	0.000	0.093	0.000	0.016	0.257	0.101	0.120	0.000	0.000	0.595	0.000	0.000	0.252	0.660	0.091	0.079
Fe^{2+}	2.606	2.441	2.483	2.355	2.574	2.136	1.781	1.887	1.902	2.540	2.072	1.842	1.819	1.897	1.074	1.537	1.617	2.134
Mg	2.275	2.115	2.366	2.376	2.242	2.695	2.870	2.902	2.921	2.254	2.789	2.618	2.253	2.217	2.869	2.455	3.231	2.725
Mn	0.039	0.036	0.052	0.078	0.000	0.058	0.026	0.027	0.087	0.043	0.000	0.041	0.000	0.024	0.036	0.000	0.050	0.047
ΣC	4.991	4.868	4.982	5.000	4.962	5.000	5.000	5.000	4.922	4.975	4.926	5.000	4.782	4.752	5.000	5.000	5.000	5.000
Ca	1.967	1.892	2.005	1.997	1.989	1.982	1.933	1.962	1.936	1.886	2.035	1.932	1.546	1.736	1.977	1.477	1.951	1.972
Na	0.033	0.108	0.000	0.000	0.011	0.018	0.067	0.038	0.064	0.114	0.000	0.000	0.454	0.264	0.000	0.298	0.049	0.028
ΣB	2.000	2.000	2.005	1.997	2.000	2.000	2.000	2.000	2.000	2.000	2.039	1.932	2.000	2.000	1.977	1.774	2.000	2.000
Na	0.179	0.220	0.000	0.000	0.086	0.076	0.001	0.056	0.047	0.094	0.000	0.000	0.176	0.242	0.000	0.000	0.079	0.088
K	0.000	0.000	0.030	0.023	0.000	0.022	0.030	0.011	0.026	0.023	0.000	0.000	0.052	0.022	0.000	0.102	0.000	0.032
ΣA	0.179	0.220	0.030	0.023	0.086	0.098	0.031	0.067	0.073	0.117	0.000	0.000	0.228	0.264	0.000	0.102	0.079	0.120

Table 6.9. Representative analyses of Epidote from the Fishguard District

	Core	Rim																
	C	C	M	M	M	M	M	M	M	M	M	M	P	P	P	P	P	P
SiO_2	36.98	37.74	38.26	37.93	37.62	37.48	37.60	37.93	37.69	37.68	38.29	37.84	39.43	38.69	37.38	38.71	38.84	39.69
TiO_2	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Al_2O_3	18.76	21.68	2.86	21.87	21.97	22.94	23.45	22.68	21.80	21.84	24.83	22.75	28.48	26.78	24.02	26.64	24.65	28.05
Fe_2O_3	17.93	14.30	9.74	14.30	15.20	13.67	13.36	12.58	14.13	13.09	11.15	13.46	5.51	7.90	5.69	6.89	10.49	5.55
MnO	n.d.	n.d	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
MgO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
CaO	23.02	23.66	23.18	24.24	23.71	23.93	24.06	23.43	23.86	23.60	24.28	24.65	24.92	24.76	20.50	24.49	24.08	24.46
Na_2O	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
K_2O	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Total	96.60	97.38	94.04	98.34	98.51	98.02	98.47	96.63	97.48	96.21	98.55	98.70	98.34	98.13	97.59	96.73	98.06	97.75
Structural formulae on the basis of 12.5 oxygens																		
Si	3.144	3.121	3.190	3.110	3.109	3.100	3.119	3.127	3.114	3.135	3.112	3.114	3.070	3.103	3.138	3.086	3.115	3.104
Al^{iv}	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
ΣZ	3.144	3.121	3.190	3.110	3.109	3.100	3.119	3.127	3.114	3.135	3.112	3.114	3.070	3.103	3.138	3.086	3.115	3.104
Al^{vi}	1.885	2.114	2.246	2.113	2.084	2.144	2.177	2.024	2.123	2.124	2.317	2.150	2.614	2.468	2.401	2.504	2.330	2.586
Ti	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.012	0.000	0.000	0.000	0.000	0.000	0.000
Fe^{3+}	1.279	0.989	0.679	0.980	1.023	0.906	0.880	0.867	0.976	0.911	0.738	0.902	0.358	0.516	0.569	0.458	0.703	0.361
Mn	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
ΣY	3.164	3.103	2.985	3.903	3.107	3.050	3.057	3.071	3.099	3.053	3.055	3.064	2.872	2.974	2.970	2.962	3.033	2.947
Ca	2.102	2.096	2.070	2.129	2.090	2.118	2.115	2.070	2.110	2.104	2.060	2.117	2.079	2.074	2.050	2.092	2.069	2.094
Mg	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
ΣW	2.102	2.096	2.070	2.129	2.090	2.118	2.115	2.070	2.110	2.104	2.060	2.117	2.079	2.074	2.050	2.092	2.069	2.094
Ps%	40.42	31.87	23.21	31.68	32.92	29.70	28.78	32.56	31.49	29.83	24.15	29.43	12.04	17.35	19.15	15.46	23.33	12.24

Table 6.8. Representative analyses of actinolite from metabasites in the Fishguard district. **Table 6.9.** Representative analysis of epiodote from metabasites (M), metabasic cataclastites (C), and pelitic cataclastites (P).



(A)
Ca
60
50
40
30
20
Al
10
20
30
40
Fe tot 3+
Pelitic shear zone
Metabasite
Ca
Al
Fe
Ca
60
50
40
30
20
Al
10
20
30
40
Fe tot 3+
FISHGUARD VOLCANIC COMPLEX
Samil Ophiolite
Merens Shear Zone
N. Peloponnesus
Franciscan Complex (Blueschist)
Prehnite-pumpellyite
Greenschist
Sedimentary Cataclastite
Epidote metadomains & metabasic cataclastites
Count
10
8
6
4
2
0
5
9
13
17
21
25
29
33
37
41
45
Ps%
(B)

and an oxidising environment throughout prograde conditions.

Local fluctuations and relative depletion of the oxygen in the fluid phase may be suggested by optically zoned epidotes observed within a cataclastite from pillow lavas at Carn Hendi. A microprobe traverse across a single zoned crystal indicated a core with an unusually high iron content (≈ 40% Ps) and a relative Al-enriched rim corresponding to a sharp optical zonation, although the rim composition was close to the theoretical $Ca_2(Fe^{3+},Al_2)Si_3O_{12}$ (OH). The reasons for unusually high Fe_2O_3 entering the epidote structure is beyond the scope of this study. Of interest however, is that it may possibly indicate that during initial fracturing and tectonism, the fO_2 (and probably a_{Fe3+}) of the precipitating fluid was extremely high, and whilst becoming depleted in time, remained sufficiently high to account for rim composition of ≈ 33 mol% Ps. Although fO_2 is likely to have been high, Liou (1973) argues that bulk composition determines epidote chemistry, whilst fO_2 determines the maximum possible Fe^{3+} content that may enter the developing crystal. In part, this appears to be substantiated by the observation of clinozoisite (9 – 14 mol% Ps) rather than epidote, crystallising within pelitic cataclastites from Porth Sychan; in intermit association with prehnite, quartz, chlorite, and neocrystic albite. The presence of prehnite and albite indicates high fO_2 {6.2.} strongly suggesting that clinozoisite is forming as a result of a high a_{Al3+}/a_{Fe3+} ratio in the fluid phase, most likely to be accountable to the host rock defining phase chemistry by the types of ions that are released into the fluid phase, rather than being defined by divariant P–T. Of interest with regards the unusually high Fe* content of the core of basaltic cataclastite epidotes, is that the depletion to relatively Al-enriched rims is also seen in prehnite from metabasic cataclastites (see Plate 6.5.). This may suggest that initial fluid which existed prior to fracturing, was either, relatively enriched in the Fe^{3+} cation and became depleted with mineral growth, or the a_{Al3+}/a_{Fe3+} increased as result of progressive restriction of circulatory fluids.

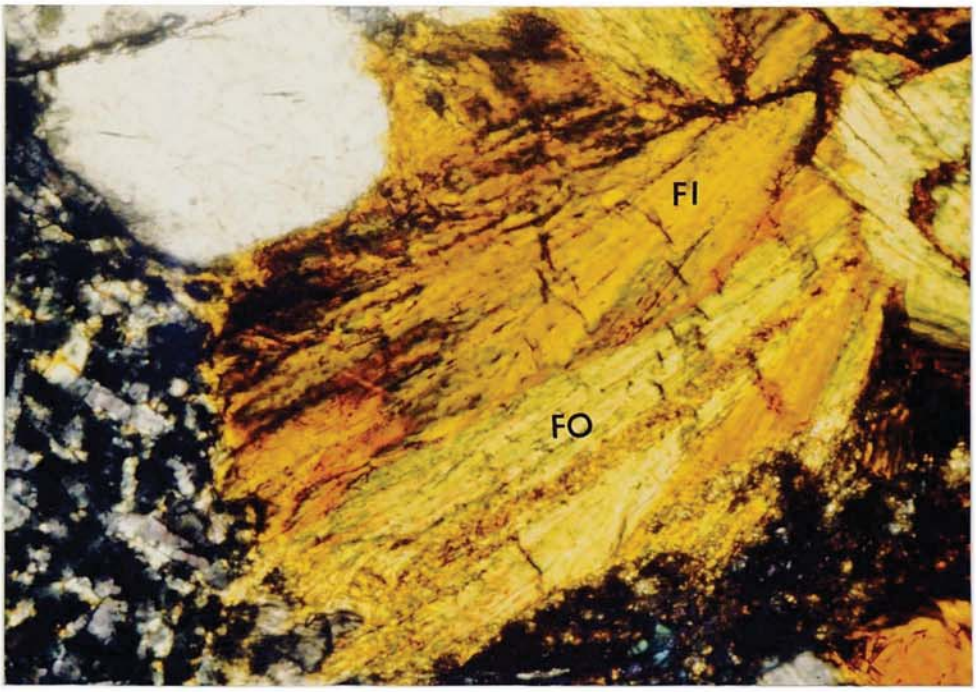
It should be noted that whist local depletion or fluctuations of oxygen in the fluid phase may be recorded by zonation and Al-enrichment trends other parameters may be similarly effective (see Bird & Helgeson 1981, Cho *et al.* 1986). For example, Cho *et al.* (1986) demonstrated from natural occurrence, a decrease in the Fe* content of epidotes during the transition from the zeolite facies to the prehnite-pumpellyite facies within the Karmutsen Volcanics, British Columbia, which they attribute to complex kinetic factors (i.e. the reverse of the Al-enrichment trend attributed to prograde conditions).

(g) Stilpnomelane

Eggleton (1972) adopts the following nomenclature: Ferrostilpnomelane, green varieties rich in ferrous iron; Ferristilpnomelane, brown to black varieties rich in ferric iron, and stilpnomelane a general term for the group as a whole. This is followed here.

Plate 6.6. Epidote from a chloritised hyaloclastite margin of a pillow lava at Pen Rhiw [S.M. 9370 3885]. Note large frambiodal aggregates of sphene (Black). Feild of veiw 0.89mm

Plate 6.7. Radiate stilpnomelane splay from a fracture zone within a doleite at Fishguard Harbour [S.M. 9590 3750]. The plate is taken in cross-polars and variation in birefringence (light to dark) reflects the development of ferro (Fo) and ferri- (Fe) stilpnomelane; the later forming through incipient oxidation of the ferro-type. The light crystal is quartz and the mottled background chlorite. Feild of veiw 0.89mm


FI
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The occurrence of stilpnomelane within the Fishguard metabasites is sporadic, although locally abundant, particularly as a component of greenschist assemblages. It is rare, although nevertheless present, in transitional and prehnite-pumpellyite assemblages. It is common to jasperiod cherts on the Pen Caer Peninsular; the inferred high iron content of which appears to be an obvious contributory factor in its development as it is absent from porcellanitic cherts. It is common in the large tonalitic intrusion that occupies the Pwl Deri headland, where it appears to be pervasively developed in a similar manner to that described by Roach (1969) from the St. Davids – Carn Lliddi intrusion, to the north of St. Davids.

In thin-section stilpnomelane is most commonly observed as poorly crystalline interstitial aggregates, and more rarely as a vein filling mineral. Its habit is variable, from amorphous to poorly crystalline aggregates (Plate 6.3.), well crystalline sheaves commonly acicular and more rarely bow-tie, and as radiating sprays and platelets (Plate 6.7.). It has been observed in both oxidation states, although ferristilpnomelane is generally dominant. Ferristilpnomelane shows a strong pleiochroism suggestive of a high Fe^{3+} (Deer *et al.* 1962), whilst interference colours are on the whole masked by its strong absorption. Ferrostilpnomelane where observed has a weak pleiochroism, exhibiting second order blue-green interference colours and low refractive index.

Based on its crystal structure, Eggleton (1972) gives stilpnomelane the mineral formula; $(Ca,Na,K)_4$ $(Ti_{0.1}Al_{2.3}Fe_{35.5}Mn_{0.8}Mg_{9.3})$ $(Si_{63}Al_9)$ $(O,OH)_{216} \cdot nH_2O$, although the ideal formula is a matter of continuing debate (Feininger 1984). Chemical analyses of stilpnomelane from a basic intrusives and a jasperiod chert are presented in Table (6.10.). Recalculation is based on normalisation of the cations Si + Al + Fe* + Mn + Mg to 7.5 the contents of the subcell; Fe computed as FeO (Eggleton 1972).

In metabasite stilpnomelane the ratio X_{Fe*} ($= n_{Fe*} / n_{Fe*} + n_{Mg}$) varies in the range 0.61 - 0.72, similar to those from some greenschist terrains (e.g. Brown 1967, Graham 1976), although less iron-rich than from others (e.g. Lopez-Montano 1984). The ratio X_{Fe*} for stilpnomelane hosted within jasperiod cherts from Maen Jaspis, is in the range 0.77 - 0.84. Whilst higher than those from metabasites such values are comparable to stilpnomelane from weakly metamorphosed Precambrian iron formations (e.g. Klien 1974, Gole 1980). An antipathetic relationship between FeO and MgO is apparent in all all analyses which may indicate $Fe^{2+} \longleftrightarrow$ Mg exchange. Disparities are evident in K_2O values between rock types although this may reflect the progressive leaching of K accompanying oxidation (see below).

The primary oxidation state of iron in stilpnomelane is problematical. It has been suggested that ferristilpnomelane forms through the oxidation of ferrostilpnomelane, commonly under atmospheric conditions. However, Eggleton (1972) argues that there is no reason to assume that all ferristilpnomelane and particularly some vein varieties formed by oxidation of ferrostilpnomelane. Sievell & Waterhouse (1984) go further by suggesting that ferro-stilpnomelane never forms in veins, indicating that all vein stilpnomelane is of ferri-type. During this study however, ferrostilpnomelane has been identified in vein and fracture systems from a small dolerite at Fishguard Harbour [S.M. 9590 3750]. Whilst it is associated with ferristilpnomelane, texturally their is little doubt as to ferristilpnomelane forming from ferrostilpnomelane as a result of incipient oxidation. The reason for this is that two thin-sections of the veins where made early in this study (one uncovered polished thin-section, one glass-slip covered thin-section); the polished thin-section contained veins of ferrostilpnomelane and small variegated patches of ferristilpnomelane, this section now contains almost total ferristilpnomelane. The covered thin-section contained two small veins, one possessing ferristilpnomelane, the other ferrostilpnomelane (which had been exposed at outcrop), both remain unchanged. The oxidation of ferrostilpnomelane at atmospheric conditions, has been observed by Sato (1975), Graham (1976) and Lopez-Montano (1984), while Frey *et al.* (1973) found that ferrostilpnomelane contained as much as ten times the amount of K_2O as ferristilpnomelane, attributable to the progressive leaching during oxidation. Brown (1971) suggests that 1.88 wt. % K_2O is the minimum content of potassium in natural occurrences of stilpnomelane which have not suffered from oxidation. The leaching accompanying oxidation may account for the disparity in K_2O content of the listed analyses (Table 6.10.). Fenninger (1984) suggested however, that K_2O content is likely to be more complex than simply the result of depletion during oxidation as he records ferrostilpnomelane from metasomatised iron-formation containing 0.6 wt. % K_2O.

Brown (1967) suggests that ferristilpnomelane should not be stable at the P-T conditions of the greenschist facies, although Nitsch (1970) demonstrated its stability under high fO_2 to conditions of the upper greenschist. More recently it has been observed from a variety of environments, including the amphibolite facies (Korikovskiy *et al.* 1975). Whilst stilpnomelane has limited use as an isograd mineral, the assemblage association of actinolite + stilpnomelane + albite + epidote + chlorite + quartz, is extremely characteristic of the greenschist facies (Turner 1981); and an assemblage that is widely developed throughout the Fishguard district.

6.4.2.(h). White Mica.

White-mica occurs in variable quantities in all rock types. In pelitic sediments it is invariably secondary, although in specific lithologies such as the *Lingula* Group {4.4.} it occurs as detrital grains. In igneous lithologies it is observed most commonly as small flakes and laths replacing feldspar, and more rarely as a ground mass phase. It may locally be common within sedimentary hosted cataclastites.

Fourteen analyses of white-mica from a metabasite and pelitic-tuffaceous sediments of the

Lower Town Formation are presented in Table 6.11. Recalculation assumes the general dioctahedral mica formula; $X_2Y_4Z_8O_{20}$ $(OH)_4$, where X= K, Na, and Ca, Y=Al, Fe, Mg, and Ti, Z= Si and Al. Recalculation is based on 22 oxygens and iron computed as Fe^{2+}, (Katagas & Panagos 1984). It is likely however, that Fe^{3+} is present in small quantities to balance the structural formula; high cation totals in the octahedral sites suggests this may be the case.

The general chemical features of the white-micas analysed are; the ratio Si:Al in the tetrahedral site is >6.5 cations per $O_{20}(OH)_4$ the ideal muscovite ratio of 3:1 is exceeded in all cases. Considerable Fe+Mg<—>Al substitution in octahedral Y-sites indicates that the white-mica is phengitic in nature. The celedonite component (Cel % = 100 Σ (Fe + Mg) / Fe + Mg + Al^{iv}) is variable from 4.9 to 21.9 %Cel in sedimentary white-micas and spans the ideal muscovite and phengite fields, whilst in metabasites the compositional range is small (20.3 to 28.9 %Cel), all analysed micas being typically phengitic in composition.

The limited number of reliable analyses, precludes a comparison between white-mica hosted within metabasites and those within sedimentary rock types; although the spread in composition of sedimentary hosted white-mica appears to be typical of white-micas from low-grade metasediments (White *et al.* 1985, Wyberecht *et al.* 1985). Metabasite white-micas compare favorably with those recorded from elsewhere in Wales (see Bevins & Rowbotham 1983), and are similar to the very low Na_2O-TiO_2 phengites recorded by Coombs *et al.* (1976) from transitional and intermediate pressure facies.

6.4.2. (I). Other phases; K-feldspar, calcite. quartz.

K-feldspar is rare, occurring as sporadic anhedral pseudomorphs after plagioclase, or as variegated patches and minor vein-fill in volcanoclastics and pillow lavas. Analyses (Table 6.2.) indicates its composition is nearly pure end-member orthoclase.

Quartz is abundant in all rock types, particularly in lavas and volcanoclastics where in may locally be chalcedonic in nature. In is invariably the dominant vein filling phase. In acidic lithologies it is difficult to distinguish quartz that resulted from primary devitrification and recrystallisation and that which resulted from subsequent metamorphism, the latter generally contains more inclusion.

Calcite is widespread, and present in at least minor amounts in all rock types. It occurs commonly in vesicules, vein, and as a replacement of feldspar. It is generally late in the paragenetic sequence, and is not considered to indicate high μCO_2 during prograde condition. It is thought rather, that its local abundance represents late stage retrogressive calcium-metasomatism, with calcium rich fluids focused along faults and fractures creating calcite metadomains.

Table 6.10. Representative analyses of Stilpnomelane

	M	M	M	M	M	M	M	M	M	M	M	M	C	C	C	C	C	C
SiO_2	46.43	48.04	47.94	49.24	45.69	49.24	50.59	46.88	47.99	47.17	46.88	47.34	46.14	46.43	48.02	51.21	46.65	45.37
TiO_2	n.d.	n.d.	n.d.	n.d.	0.29	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Al_2O_3	5.41	6.21	5.63	5.26	7.22	5.96	5.63	5.65	5.88	5.56	5.44	5.52	4.70	5.08	4.40	4.12	4.49	4.53
FeO*	27.61	25.63	25.10	24.50	25.89	25.49	26.19	24.64	26.37	24.87	24.82	24.82	29.16	31.12	29.57	29.44	31.53	31.65
MnO	0.78	0.92	0.91	0.78	0.99	0.98	0.75	0.72	0.74	0.84	0.66	1.02	n.d.	n.d.	0.26	0.27	n.d.	0.23
MgO	5.99	8.99	8.50	8.64	8.78	8.72	8.91	8.37	7.89	8.54	8.73	8.65	3.13	3.31	3.60	3.26	5.32	5.28
CaO	0.18	0.25	0.26	1.59	0.23	0.19	1.35	0.22	0.22	0.17	n.d.	0.20	n.d.	n.d.	n.d.	n.d.	0.23	0.53
Na_2O	0.47	n.d.	n.d.	n.d.	n.d.	n.d.	0.42	n.d.	n.d.	n.d.	0.38	n.d.	0.83	0.85	0.81	0.53	0.52	0.58
K_2O	0.34	0.47	0.24	0.42	0.48	0.37	0.53	0.30	0.49	0.16	3.77	0.69	2.43	2.70	2.25	2.02	3.41	3.12
Total	87.21	90.51	88.58	90.43	89.57	90.95	94.37	86.78	89.58	87.31	90.68	88.24	86.39	89.49	88.91	90.85	92.15	91.29
Recalculation based on 7.5 cations in the subcell ($\Sigma Z + \Sigma Y$)																		
Si	4.060	3.953	4.034	4.127	3.781	4.037	4.074	4.027	4.028	4.020	4.015	4.021	4.282	4.174	4.307	4.475	4.053	4.001
Al^{iv}	0.000	0.047	0.000	0.000	0.219	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
ΣZ	4.060	4.000	4.034	4.127	4.000	4.037	4.074	4.027	4.028	4.020	4.015	4.021	4.282	4.174	4.307	4.475	4.053	4.001
Al^{vi}	0.558	0.555	0.558	0.519	0.486	0.576	0.534	0.572	0.581	0.559	0.549	0.553	0.514	0.538	0.465	0.424	0.460	0.472
Ti	0.000	0.000	0.000	0.000	0.219	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Fe^{3+}	2.021	1.764	1.766	1.717	1.794	1.748	1.764	1.770	1.851	1.772	1.772	1.763	2.264	2.339	2.219	2.152	2.219	2.335
Mn	0.057	0.064	0.065	0.055	0.069	0.068	0.051	0.052	0.053	0.061	0.048	0.073	0.000	0.000	0.019	0.020	0.000	0.019
Mg	0.782	1.103	1.067	1.079	1.084	1.065	1.069	1.072	0.988	1.086	1.114	1.095	0.434	0.444	0.476	0.425	0.690	0.694
ΣY	3.418	3.486	3.456	3.370	3.461	3.457	3.418	3.466	3.473	3.478	3.489	3.495	3.312	3.321	3.179	3.019	3.441	3.520
Ca	0.015	0.021	0.022	0.131	0.194	0.016	0.107	0.018	0.019	0.014	0.000	0.000	0.015	0.000	0.000	0.000	0.020	0.046
Na	0.069	0.000	0.000	0.000	0.000	0.000	0.065	0.000	0.000	0.000	0.054	0.000	0.135	0.135	0.126	0.080	0.080	0.091
K	0.188	0.067	0.045	0.172	0.242	0.043	0.173	0.049	0.068	0.031	0.377	0.069	0.256	0.227	0.231	0.201	0.346	0.321
XFe*	0.72	0.61	0.62	0.63	0.62	0.62	0.62	0.62	0.65	0.62	0.61	0.62	0.84	0.84	0.82	0.83	0.77	0.77

Table 6.11. White mica analyses.

	M	M	M	M	M	M	S	S	S	S	S	S	S	S
SiO_2	46.97	49.77	48.28	48.52	48.12	45.74	45.41	48.61	46.42	48.40	48.40	45.98	46.51	46.49
TiO_2	0.32	0.49	0.40	0.47	0.37	0.46	n.d.	n.d.	n.d.	n.d.	n.d.	0.35	0.20	0.29
Al_2O_3	32.96	34.84	34.48	34.19	34.43	32.65	30.06	29.46	30.81	34.00	34.00	33.29	33.92	35.48
FeO*	3.05	0.93	0.87	1.19	0.83	5.14	6.44	4.76	5.31	1.02	1.02	3.16	1.63	2.38
MnO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
MgO	1.18	0.72	0.57	0.61	0.73	1.77	2.23	1.57	1.78	0.39	0.39	0.91	0.51	0.38
CaO	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.	n.d.
Na_2O	n.d.	n.d.	0.49	n.d.	0.31	0.38	0.54	n.d.	0.71	n.d.	n.d.	0.31	0.32	1.38
K_2O	9.09	9.56	9.77	9.69	9.41	8.55	9.37	8.89	9.74	9.42	9.41	8.59	9.43	9.05
Total	93.57	96.31	94.86	94.67	94.20	94.69	94.05	93.29	94.77	93.24	93.23	.92.59	92.52	95.45
Structural formulae based on 14 oxygens														
Si	6.347	6.447	6.382	6.442	6.391	6.186	6.270	6.630	6.330	6.478	6.478	6.278	6.325	6.194
Al^{iv}	1.653	1.553	1.618	1.578	1.608	1.914	1.730	1.370	1.670	1.522	1.522	1.722	1.675	1.806
ΣZ	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000	8.000
Al^{vi}	3.597	3.767	3.754	3.755	3.772	3.291	2.965	3.366	3.283	3.842	3.842	3.635	3.763	3.712
Fe	0.344	0.101	0.097	0.124	0.093	0.581	0.745	0.543	0.606	0.115	0.115	0.361	0.185	0.263
Ti	0.033	0.048	0.040	0.047	0.093	0.047	0.000	0.000	0.000	0.000	0.000	0.036	0.021	0.000
Mg	0.237	0.140	0.113	0.120	0.145	0.357	0.459	0.319	0.362	0.079	0.079	0.185	0.104	0.076
ΣY	4.211	4.056	4.004	4.046	4.103	4.276	4.169	4.228	4.251	4.026	4.062	4.217	4.073	4.051
Ca	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
Na	0.000	0.000	0.127	0.000	0.080	0.100	0.145	0.000	0.188	0.000	0.000	0.084	0.085	0.355
K	1.567	1.580	1.648	1.637	1.592	1.475	1.652	1.547	1.695	1.608	1.608	1.497	1.636	1.523
Cel %	13.91	6.01	5.29	6.10	5.94	21.93	28.99	20.38	22.77	4.09	4.80	13.0	7.09	8.36

Table 6.10. Representative analyses of stilpnomelane from metabasites (M) and a jasperiod chert (C). **Table 6.11.** Representative analyses of white-mica from metabasites (M) and pelitic sediments (P).

6.5. ASSESSMENT OF EQUILIBRIA

There is considerable evidence within the Fishguard metabasites to indicate disequilibrium on the scale of a thin-section. The presence of clinopyroxene in a relatively unaltered state, and plagioclase possessing a variable anorthite component, indicates that these phases have not fully reacted under the P-T-fluid conditions imposed during metamorphism. Disequilibria is also indicated by the presence of incomplete reactions such as; hornblende-actinolite pairs, and corroded aggregates of prehnite and pumpellyite in transitional assemblages. Preferential phase development in primary sites also indicates partial disequilibrium; for example, epitaxial overgrowths and replacement of clinopyroxene by actinolite under high X_{CO2} (i.e. $5CaMgSi_2O_6 + H_2O + 3CO_2 \longleftrightarrow Ca_2Mg_5Si_8O_{22} (OH)_2 + 3CaCO_3 + 2SiO_2$) and does not represent an actinolite isograd as the reactants do not form a stable assemblage under the conditions of metamorphism (Hashimoto 1972). It should be noted however, that the smaller the scale of which equilibria is sought the likely inference of disequilibria diminishes, e.g. greenschist sub-assemblage of 'transitional' assemblages are in textural equilibrium over one millimetre (Plate 6.2.), although disequilibria is prevalent on the scale of a thin-section.

It has been suggested that an approach to equilibrium in low-grade metabasites is indicated if the phase rule is obeyed. However, whilst compliance to the phase rule may be highly suggestive in that it accords well with the general thermodynamic prediction, its application falls short of proving equilibrium (Turner 1981). It has been suggested that the phase rule is invariably obeyed in the metabasites of the Welsh Basin (Bevins & Rowbotham 1983). This is substantiated in part from the Fishguard district, if one assumes that the system can be discussed in terms of the major chemical components H_2O-SiO_2-TiO_2-CaO-Al_2O_3-MgO-Na_2O- K_2O-FeO-Fe_2O_3, and that the fluid phase is effectively under external control (Smith *et al.* 1982); with SiO_2, TiO_2 and Na_2O being fixed in quartz, sphene, and albite respectively and K_2O being accounted for by white-mica in the general absence of K-feldspar. Thus leaving four variable components (on the assumption that FeO and MgO as isomorphous), allowing a four phase assemblage of actinolite, chlorite, epidote, prehnite and pumpellyite to coexist. Metabasite assemblages generally comply with the phase rule, although transitional assemblages {6.2.1.} may also obey the phase rule suggesting its applicability is limited as disequilibria can be inferred from textural relationships. However, transitional assemblages are of interest, in that whilst the phase rule may or may not be obeyed and textural disequilibrium is apparent, the corroded and variegated nature of prehnite and pumpellyite suggests that equilibria attainment was being approached; slow reaction rates may account for their relic persistence. An approach to equilibrium is also indicated by the limited intergranular chemical variation and the rarity of optical and chemical zonation in most phases (excluding cataclastites). The latter may however, be more apparent than real, due to the fine grain size of many of the phases analysed.

6.6. PHASE RELATIONS AND CONDITIONS OF METAMORPHISM

In the light of the above evidence for disequilibrium prevailing locally, the following discussion on mineral paragenesis and conditions of metamorphism is presented with obvious caution. It is also noted that the quantitative P-T values derived at, are based on experimental phase relations which are commonly developed for a specific set of assumptions, deviation from which can significantly modify relationships (Kay 1984). Nevertheless, careful application can provide a perspective on the likely paragenesis and valuable information on the metamorphic conditions which prevailed during regional metamorphism of the Fishguard district.

Phase relations

Utilizing both experimental phase relations and natural paragenesis, phase equilibria in low-grade metabasites has been studied by many. Probably the most definitive is the recent study by Liou *et al.* (1985) who discuss relationships in the model basaltic system NCMASH (Na_2O-CaO-MgO-Al_2O_3-SiO_2-H_2O) constructing petrogenetic grids and defining facies boundaries by key reactions (Fig. 6.9.a.). They also consider the effect of introducing an Fe_2O_3 component into the model system, which causes the systematic P-T displacement of both univariant lines and invariant points, designating displacements as a series of $X^{Ep/Fe}$ isopleths (= Ps component of epidote, Fig. 6.9.b.).

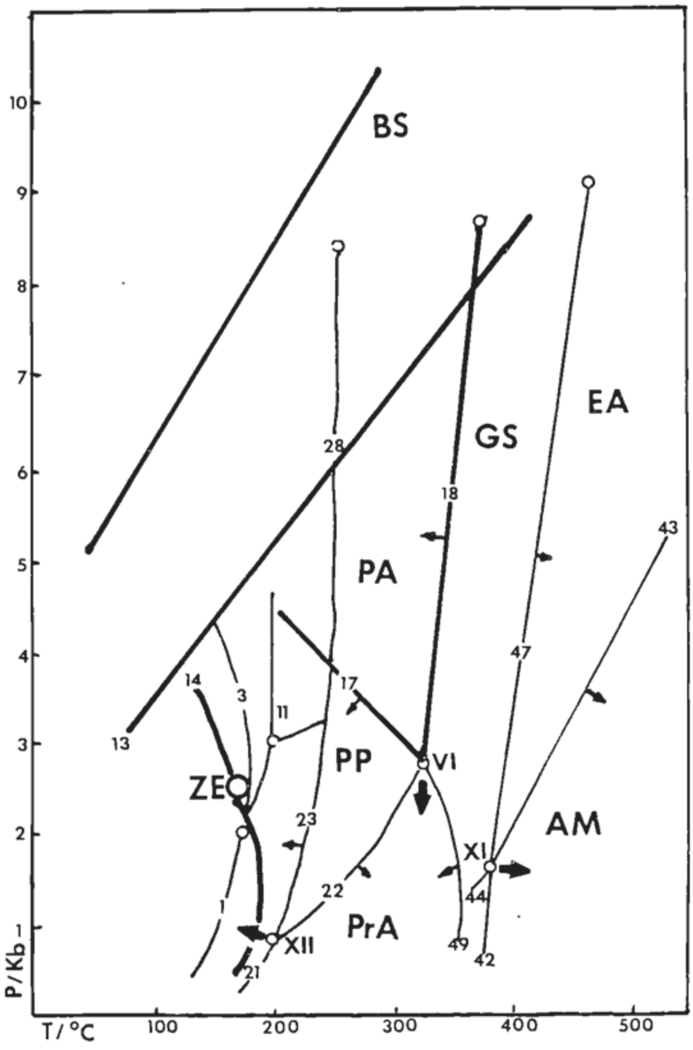
For the purpose of discussing mineral paragenesis and metamorphic conditions, it is convenient although possibly misleading, to assume that phase development is the result of prograde conditions under the singular influence of normal burial. Based on this assumption, it is likely that the observed prehnite-pumpellyite assemblages (excluding metadomain assemblages {6.3.}) owe, in part, their development to a precursor zeolitic mineralogy; the absence of which indicates that the conditions of the zeolite facies was surpassed. It is plausible that specific styles of alteration (such as the widespread prehnitisation of hyaloclastites) owe their assemblage development to progressive burial, such lithotypes being favoured by initial zeolitic growth by virtue of their high comparative permeability (Fisher & Schmincke 1984). The frequent occurrence of quartz in prehnitised hyaloclastites may result from dehydration reactions such as;

$$CaAl_2Si_4O_{12} \cdot 4H_2O + 2(OH)^- = Ca_2Al_2Si_3O_{10}(OH)_2 + SiO_2 + H_2O$$

(i.e., laumontite = prehnite + quartz + H_2O, *c.f.* Houghton 1982).

Figure. 6.9.a. P_{fluid} — T diagram after Liou *et al.* (1985), showing a petrogenetic grid for various low-grade metamorphic assemblages in the model basaltic system (NCMASH, + quartz+albite+chlorite) with selected univariant lines and invariant points. Displacement of reaction curves and invariant points due to the introduction of Fe_2O_3 are shown by arrows. Letters refer to metamorphic facies (i.e. ZEO = Zeolite facies, PP = Prehnite-Pumpellyite facies; GS = Greenschist facies; PrA = Prehnite- actinolite facies; PA = Pumpellyite-actinolite facies; EA = Epidote- Amphibolite facies: AM = Amphibolite Facies; BS = Blueschist facies). Numbers refer to reactions quoted in Liou *et al.* (1985); (i.e. 14 Analcime + Quartz = Albite + Fluid; 17 Prehnite + Chlorite + Fluid = Pumpellyite + Tremolite + Quartz; 21 Laumontite + Prehnite = Zoisite + Quartz + Fluid; 22 Pumpellyite + Quartz = Zoisite + Prehnite + Chlorite + Fluid; 23 Laumontite + Pumpellyite = Zoisite + Chlorite + Quartz + Fluid; 42 Oligoclase + Tremolite + Chlorite = Hornblende + Albite + Fluid; 43 Zoisite + Chlorite + Albite + Quartz = Oligoclase + Hornblende + Fluid; 44 Zoisite + Chlorite + Albite + Quartz = Oligoclase + Tremolite + Fluid; 47 Zoisite + Chlorite + Tremolite + Quartz = Hornblende + Fluid; 49 Prehnite + Chlorite + Quartz = Zoisite + Tremolite + Fluid).

Figure. 6.9.b. P_{fluid} — T diagram after Liou *et al.* (1985), showing selected continuous reactions around invariant points VI and XII from Figure (6.9.a.) and displacement of the invariant point along a discontinuous reaction at the introduction of Fe_2O_3 into the model basaltic system; with a series of isopleths as X_{Fe} [= $Fe^{3+}/(Fe^{3+} + Al)$] of epidote for the continuous reactions. Epidotes from the Fishguard metabasites have an approximate X_{Fe} of 0.30. Key; Pr = Prehnite, Pm = Pumpellyite, Ep = Epidote, Chl = Chlorite, Tr = Tremolite, Lm = Laumontite. PrA = Prehnite-actinolite facies, PP = Prehnite-pumpellyite facies, GS = Greenschist facies, PA = Pumpellyite-Actinolite facies, ZOE = Zeolite facies.



BS
GS
EA
PA
PP
ZE
AM
PrA
VI
XI
XII
P/Kb
T/°C
100
200
300
400
500
1
2
3
4
5
6
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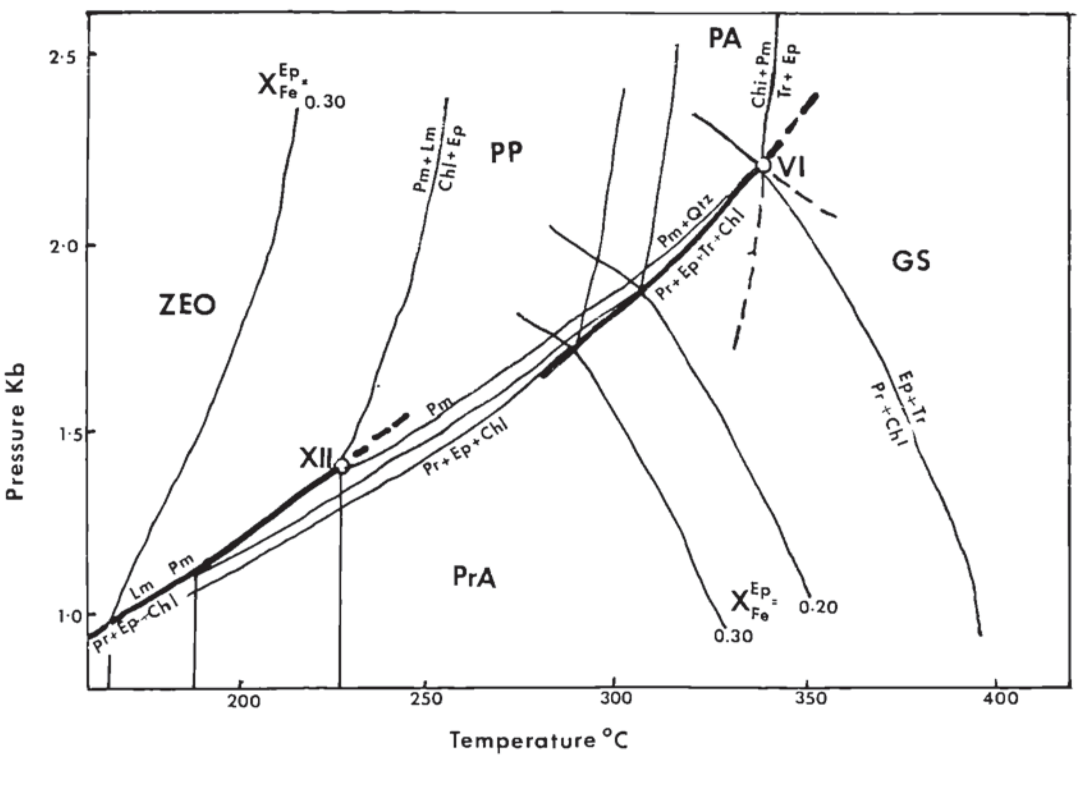


PA
PP
ZEO
GS
PrA
VI
XII
X_{Fe}^{Ep} 0.30
X_{Fe}^{Ep}= 0.30
0.20
Pm + Lm
Chl + Ep
Chl + Pm
Tr + Ep
Pm + Qtz
Pr + Ep + Tr + Chl
Pm
Pr + Ep + Chl
Lm
Pm
Pr + Ep + Chl
Ep + Tr
Pr + Chl
Pressure Kb
Temperature °C
1.0
1.5
2.0
2.5
200
250
300
350
400

The transition from the zeolite facies to the prehnite-pumpellyite facies, which can be considered the minimum temperature of metamorphism in the Fishguard district, has been determined experimentally by Liou *et al.* (1985) using the reaction laumontite + pumpellyite = clinozoisite + chlorite + quartz + H_2O at P = 0.9 - 3.1 kbars and T = 200 - 230°C, (Fig. 6.17). More recently from natural paragenesis, Cho *et al.* (1986) discuss the zeolite facies – prehnite-pumpellyite facies transition in terms of the key reaction; laumontite + pumpellyite + quartz = prehnite + epidote + chlorite + H_2O, calculating the reaction equilibrium as of P = 1.1 ± 0.5 kbars at T = 190 ± 30°C. This computes to a geothermal gradient in the range 40 – 50 °C km^{-1}, a heat flow comparable to that calculated to have existed in the Fishguard district (see below). However, Frost (1980) regarded the zeolite facies to prehnite-pumpellyite facies transition in the system CaO-Al_2O_3-SiO_2-H_2O-CO_2 as multivariant, suggesting that compositional factors such as Na content as well as FeO and Fe_2O_3 may have as great an effect as does temperature on the facies transition.

During metamorphism, low pressures (<2–2.5 kbars) can be calculated quantitatively from phase relations (with additional supportive evidence such as; the absence of high pressure phases, the lack of pumpellyite-actinolite equilibrium assemblages, the presence of Fe-enriched pumpellyite and epidote). Whilst pumpellyite is associated with actinolite in transitional assemblages, it is always observed in textural disequilibria, there is no evidence to indicate mutual coexistence or relationship by involvement in reaction such as;

calcite + epidote + chlorite + quartz = pumpellyite + actinolite + H_2O + CO_2

the calculated reaction equilibria of this reaction being T = 330 - 345°C at P = 4 - 6 kbars (Brown 1977), or the key reaction,

prehnite + chlorite + quartz = pumpellyite + actinolite + H_2O

used by Liou *et al.* (1985) to define experimentally the prehnite-pumpellyite to pumpellyite-actinolite facies transition (Fig. 6.9.), the reaction equilibria occurring between T = 230 - 320°C and P = 3.75 - 2.75 kbars. However, Liou *et al.* (1985) point out that there is no experimental or theoretical grounds for invoking a minimum pressure limit for the pumpellyite-actinolite facies. Bevins & Rowbotham (1983) record pumpellyite-actinolite assemblages from the Prescelly Hills and dolerites east of Fishguard, although the assemblage that they list contains pumpellyite + actinolite + prehnite + epidote; it being likely that such an assemblage equates with the assemblages referred to here as 'transitional' {6.2}, rather than the pumpellyite-actinolite facies. Such an inference is based on the low pressure P-T path thought most likely to suit the assemblages in the Fishguard district, and the documented occurrence of the pumpellyite-actinolite facies assemblage to deeply buried volcanics (Coombs *et al.* 1970, Schermerhorn 1975, Baltazis & Katagas 1984).

Prehnite and actinolite may similarly coexist in transitional assemblages, although textural disequilibria precludes the designation of the prehnite-actinolite facies. It is of interest to note however, the relative relic persistence of prehnite against pumpellyite in such assemblages. While this may represent an original modal predominance, it may also indicate that the 'pumpellyite-out' isograd occurred prior to the 'prehnite-out' isograd. If the latter is the case, it may not be totally unexpected, as the envisaged P-T path skirts the upper-P-boundary of the prehnite-actinolite facies, the definition of which is also supportive in that it is defined as: a low pressure facies series characterised by the disappearance of pumpellyite and the appearance of actinolite at P-T conditions transitional from the prehnite-pumpellyite facies to the greenschist facies (*c.f.* Liou *et al.* 1985).

Textural evidence within transitional assemblages indicates that prehnite and pumpellyite are in disequilibria, and breaking down to form actinolite and epidote. This may suggest that greenschist assemblages owe in part their development under progressive conditions to reactions such as;

$$32\ Ca_2Al_2Si_3O_{10}\ (OH)_2 + 5\ Mg_4Al_4Si_2O_{10}\ (OH)_8 + 10\ SiO_2 \longleftrightarrow$$

$$28\ Ca_2Al_3Si_3O_{12}\ (OH) + 4\ Ca_2Mg_5Si_8O_{22}\ (OH)_2 + 34\ H_2O$$

(prehnite + chlorite + quartz = epidote + actinolite + H_2O)

$$9\ Ca_4Al_5MgSi_6O_{21}\ (OH)_7 + Mg_6\ Si_4O_{10}\ (OH)_8 + 11\ SiO_2 \longleftrightarrow$$

$$3\ Ca_2Mg_5Si_8O_{22}\ (OH)_2 + 15\ Ca_2Al_3Si_3O_{12}\ (OH) + 25\ H_2O$$

(pumpellyite + antig. mol. of chlorite + quartz = tremolite + clinozoisite + H_2O)

Whilst many greenschist assemblages may have developed by the above reactions, evidence for actinolite + epidote + stilpnomelane assemblages in close proximity to calcite + quartz veins may suggest that reactions such as;

$$3\ Mg_3Al_2Si_3O_{10}\ (OH)_8 + 10\ CaCO_3 + 21\ SiO_2 =$$

$$3\ Ca_2Mg_5Si_8O_{22}\ (OH)_2 + 2\ Ca_2Al_3Si_3O_{12}\ (OH) + 8\ H_2O + 10\ CO_2$$

(chlorite + calcite + quartz = actinolite + epidote + H_2O + CO_2)

were locally operative in producing greenschist assemblages. The absence of hornblende, lack of evidence for a peristerite gap, or an increase in the An content of plagioclase within greenschist facies assemblages, would indicate that conditions of the amphibolite facies were not attained (i.e. $T > 380°C$, Fig. 6.9.).

Utilising the P-T-$X^{Ep/Fe}$ diagram (Fig. 6.9.), it is possible to quantitatively assess the conditions of metamorphism in the Fishguard metabasites. The pistacite component of epidotes analysed, shows no distinctive evidence for change between facies types, the majority of analysed epidotes having a Ps% close to the theoretical 33% end member {6.4.2.}, and an average $X^{Ep/Fe} = 0.30$ is therefore realistically assumed.

Using the P-T-$X^{Ep/Fe}$ diagram (Fig. 6.9.b.), the observed prehnite-pumpellyite assemblages may have formed at conditions in the range T = 160 - 290°C between 1 - 2 kbars. Temperatures are likely to have been in excess of 200 - 230°C, as true chlorites, rather than mixed layer smectite/chlorites are developed (Hoffmann & Hower 1979). Low pressure is consistent with the absence of high pressure phases, pumpellyite - actinolite assemblages, estimated stratigraphic overburden and pumpellyite chemistry. Given the assumption of a uniform heat flow under an envisaged geothermal gradient in the range 40-45°C km^{-1} (see below), then a pressure in the order of 1.2 - 1.7 kbars would be appropriate for the observed prehnite-pumpellyite facies assemblages.

As one passes south from the Pen Caer peninsular, prehnite-pumpellyite assemblages appear to give way to assemblages of 'transitional' type (Fig. 6.1.). Such assemblages are of specific interest, in that whilst reactions have clearly not gone to completion and disequilibrium can be inferred, it is likely that they represent phase development under conditions close to the sub-greenschist - greenschist facies boundary. Under low pressures in the model system NCMASH, this transition occurs at ≈T = 340°C and P = 2.5 kbars (invariant point VI, Fig. 6.9.). Introduction of Fe_2O_3 into the system displaces the invariant point along a discontinuous reaction, and where $X^{Ep/Fe} = 0.30$, boundary conditions for the continuous reactions occur at T = 290°C and P = 1.70 kbars. It is thought that such conditions may be appropriate for the transition from the prehnite-pumpellyite to greenschist facies within the Fishguard metabasites.

The lower-T-limit for development of the observed greenschist assemblages is essentially defined by the conditions of transition (i.e. T > 290°C), and its upper-T-limit by the lower boundary of the amphibolite facies (i.e. T ≤ 380°C, Fig. 6.9.). It has been suggested that an approach to the amphibolite facies in low pressure regimes may be indicated by the development of actinolite + oligoclase (Liou & Ernst 1979) and two - sodic/calcic plagioclase + Ca-amphibole assemblages (Maruyama *et al.* 1983). The absence of An enrichment or a peristerite gap, possibly indicates that the observed greenschist assemblages developed at temperatures towards conditions of the lower greenschist facies boundary (T < 350°C) and at pressures possibly less than 2 kbars.

Non-divariant P–T factors: Whilst mineral paragenesis and assemblage development is appropriately discussed in terms of divariant P–T, other factors are likely to have been important in governing phase development and mineral chemistry within the Fishguard metabasites. The role of a fluid phase appears to be essential, with variation in the fluid chemistry, activity of ions, oxygen fugacity, permeability (original and tectonic induced) and precursor mineralogy being factors likely to have contributed to development of the observed assemblages.

Textural evidence indicates alteration is integrally related to reaction with a fluid. The intense alteration of hyaloclastites and fragmentary volcanoclastics, and marked decrease in the degree of alteration within competent impermeable strata, strongly suggests a fluid-dominated system. Areas of high permeability exhibit pervasive alteration and recrystallisation, whilst tectonism clearly created conduits though which fluids passed, facilitating vigorous alteration and rapid phase development, particularly in close proximity to the site of deformation.

Studies of element mobility during low-grade metamorphism, suggests that many are susceptible to alteration by secondary processes (e.g. Smith & Smith 1976, Humphris & Thompson 1978, Hynes 1986). Bevins (1982), in a detailed chemical and petrogenetic study of the Fishguard Volcanic Complex, suggests that whilst some elements may show loss or gain, many appear to have been isochemical. The presence of polymineralic vesicular infilling and polymineralic assemblages within cataclastites, attests to at least microdomain scale mobility, and possible regional scale mobility of the majority of major elements and probably a significant proportion of trace elements in a fluid phase. It follows, therefore, that phase development and chemistry is likely to be defined in part by the type of ions released into the fluid, as well as conditions of P-T; determined primarily by the igneous phases undergoing alteration, their susceptibility to brakedown, pH and temperature. The general lack of optical and chemical zonation and limited intergranular and inter-specimen variation, suggests a_{Ca}, a_{Al}, a_{Si}, a_{Fe2+}, a_{Fe3+}, a_{Mg}, a_{Na}, and a_{Ti} remain high and relatively constant, as fluctuations in time tend to lead to chemically heterogeneous phases (Offler 1984). The role of precursor mineral chemistry in defining the chemistry of replacing phases appears to be limited, as indicated by such replacements as plagioclase by pumpellyite and epidote, requiring Fe-Mg introduction into the feldspar site.

The availability of specific cations does not mean that excessive quantities of that cation will enter the developing phase, as variables of P–T, fO_2 and mineral buffers are likely to have controlled cation content. For example, Fe-pumpellyite is commonly developed in low pressure environments whilst its Al counterpart is generally restricted to high pressure terrains {6.4.4.}, commonly attributed to declining fO_2 under progressive conditions, a phenomenon suggested also suggested as the major influence in crystallising zoned epidotes in low grade terrains (i.e. Fe-rich cores, Al-rich rims). Crystallisation under high fO_2 may also favour amphiboles enriched in Fe^{3+}, Mg, Si and Mn (Spears 1981), whilst Ti substitution in the amphibole C-site is favoured by an increase in temperature and a decrease in pressure (Hynes 1982).

Within the Fishguard metabasites, the Fe-rich nature of pumpellyite and a dominance of epidote with a high pistacite component, suggest that during prograde conditions the environment was essentially oxidising with high fO_2 readily fixing ferric iron in silicates (it being likely that the oxygen fugacity was higher than the FMQ buffer and possibly as high as the HM buffer which produces epidote with a maximum Fe^{3+} content; Sievell & Waterhouse 1984). The general absence of zoned epidotes, or enrichment in the Al content of epidotes from the various assemblages and defined facies, suggests that fO_2 remained high in time. A possible explanation as to fO_2 remaining high during prograde conditions, is that circulation of the metamorphic fluid was facilitated by an original high permeability, accentuated by subsequent tectonism. This would permit the circulation of oxygen-rich fluids, accounting for a predominance of epidote–haematite–sphene assemblages (Evarts & Schiffman 1983), and relative scarcity of actinolite. Low μCO_2 during prograde conditions, is attested by the presence of prehnite, albite and widespread persistence of sphene (i.e. Schuiling & Vink 1967, Thompson 1971, Evarts & Schiffman 1983). Late within the paragenetic sequence however, possibly under retrogressive conditions, a reducing environment is suggested by the local development of ferrostilpnomelane – calcite veinlets, and widespread development of large calcite metadomains and veins. This may suggest that permeability was decreasing, whilst the progressive removal of H_2O resulting from hydrous phase development, facilitated a decrease in the activity of H_2O, increasing a_{CO2}/a_{H2O} sufficiently to precipitate calcite (Thompson 1971).

Potential Lower Palaeozoic overburden: The calculated P–T (T = 230-350°C, P = 1.2-2.0 kbars) for development of the observed assemblages, assuming normal burial conditions and a load pressure of 250 bars per km^{-1}, implies a geothermal gradient approximating 40–45°C km^{-1}; requiring a stratigraphic overburden in the range 6–8 km to produce the observed assemblages. A geothermal gradient of 40–45°C km^{-1} appears high, given an average background geothermal gradient of 25–30°C km^{-1}, although comparable to that visualised by Robinson & Bevins (1986) of ≈42°Ckm^{-1} to have been regional across the Welsh Basin {6.7.}.

As recently demonstrated by Alderidge (1986), with regards to metamorphism along the SE margin of the Welsh Basin, a test of metamorphic conditions can be undertaken by an assessment of potential Lower Palaeozoic overburden. The presence of tectono-metadomains indicates that the envisaged metamorphic conditions were prevailing at the time of Caledonide deformation (i.e. end Silurian – early Devonian, see Chapter 3), strongly suggesting that any subsequent deposition above the Old Red Sandstone is unlikely to have contributed to the regional metamorphic geotherm. The lack of Ashgillian, Silurian and Devonian strata within the immediate vicinity of the Fishguard district, requires the supposition that the thickness of such strata around the Haverfordwest district and throughout South Dyfed, is indicative of potential cover across S.W. Dyfed (see, George 1970, Cocks *et al.* 1971, House *et al.* 1977); although it should be noted that throughout much of this area, thickness were controlled in the most part by the reactivation of Caledonide structures resulting in the formation of basin and horst geometries (Dunne 1980).

Noting the likely possibility of diachronous deposition between south and south-west Dyfed, arguably, a stratigraphic cover above the Fishguard Volcanic Complex of post-Caradoc – pre-Upper Old Red Sandstone in the range 4.2 – 5.5 kms is plausible, although 5.5 km may be excessive. Llandeilo and Caradoc sediments in the vicinity of Dinas Head, east of Fishguard, have an estimated outcrop thickness of between 1.0 – 1.4 kms. The likely Upper Llanvirn age of the Fishguard Volcanic Complex and an eruptive thickness in excess of 1.5 kms (Chapter 5), indicates that prior to tectonism, the stratigraphic overburden above the base of the volcanic complex is likely to have been in the region of 6.7 – 8.4 kms (≈ 1.6 – 2.1 kbars), whilst the upper most extrusives are likely to have been buried to depths of 5.2 – 6.8 kms (≈1.3 – 1.7 kbars). Values for stratigraphic cover may be regarded at best as creditable estimates, they coincide well with the envisaged pressures and an geothermal gradient of 40 – 45°C calculated from the observed assemblages (i.e. prehnite-pumpellyite facies assemblages, T= 230 – 290°C at P= 1.2 – 1.7 kbars; greenschist facies assemblages T= 290 – 350°C at P= 1.7 – 2.0 kbars).

It seems unlikely that the observed assemblages, calculated P–T, and estimated stratigraphic overburden are coincidental; it being suggested that conditions of metamorphism are accountable in part to the operative process of burial under an unusually high geothermal gradient. A conclusion which concurs with that recently drawn by Robinson & Bevins (1986) with regards to metamorphism across the Welsh Basin.

6.7. STYLES OF ALTERATION, COMPARATIVE STUDIES & SUMMARY.

Early alteration

A sequence of events responsible for observed styles of alteration and metamorphism, begins with early post-eruptive/intrusive alteration. Extrusives are likely to have been susceptible to rapid alteration and chemical change at surfacial levels, resulting from abundant sideromelane and permeable volcanoclastics and hydroclastics. During early alteration, P_{H2O} may approximate hydrostatic pressure and not lithostatic pressure, allowing relatively free circulation of fluids (Houghton 1982). The rates of alteration are likely to be governed by variable water/rock ratios, pH, temperature and permeability, with high fluid rock ratios ensuring fluid compositions were buffered by host rock chemistry (Sievell 1984). The occurrence within high-level gabbroic intrusions of a hornblende ± quartz assemblage (subsequently retrogressed by low-grade regional metamorphism), indicates that early alteration was not restricted to extrusive lithologies; although the localised nature of such assemblages possibly indicates lower permeability and restricted transfer of reacting fluids. Deuteric reactions with magmatic fluids are likely to have occurred, whilst hydrothermal circulation of seawater permeating to shallow levels from surfacial fractures and permeable strata is also highly probable. Whilst conjectural, 'patch–like' epidote metadomains common to many larger gabbroic bodies {6.2.} may represent a continuum under falling temperatures, of early post-magmatic alteration as fluids within the gabbroic bodies cooled, analogous to hydrothermal alteration associated of oceanic basalts (e.g. Mevel 1982, Liou & Ernst 1979).

During active volcanism and emplacement of high-level contemporaneous intrusives, it is likely that high anomalous heat flow in comparison to the background geotherm would have been prevalent throughout the Fishguard distict and much of SW Dyfed. Given the volcanotectonic environment envisaged for the Welsh Basin during the Ordovician, of an extensional marginal basin founded on immature and possibly thinned continental crust {1.1}, by modern day analogy, a background geothermal gradient of ≈ 25°C km^{-1} (≈ 50 mWm^{-2}) appears inappropriate. Kokelaar *et al.* (1984a), express the opinion that whilst no single modern day analogy of the Welsh Basin exists, the extensional regime and environmental setting of the Taupo Volcanic Zone and Rotorua Depression, New Zealand, show marked similarities; within which Cole (1985) records heat flow values up to 27 times higher (i.e. 825–1575 mWm^{-2}) than normal continental crust. Clearly such a heat flow is singularly inappropriate in the context of regional metamorphism of the Welsh Basin, although serves as an example for highly anomalous heat flow in extensional marginal basin terrains. In modern back-arc basins with few exceptions, heat-flow approximates to 1.5 - 3.0 times higher than the 'mean' values of tectonically stable regions (i.e. Andrews & Sleep 1974, Sychev & Sharaskin 1985), such heat flow accounting for 'baric' type metamorphism (low P – high T). It seems likely, given the widespread development of igneous activity and extensional regime of the Welsh Basin during the Ordovician, that a mean geothermal gradient of +35 °C km^{-1} is realistic, whilst in areas of active volcanotectonism a geotherm possibly as high as 60°C km^{-1} may have been appropriate.

Conditions of metamorphism and comparison with other studies

Following early alteration in the Fishguard metabasites, it is thought that the physical conditions of 'regional' metamorphism would have been comparable in many respects to that recorded from elsewhere in the SW Dyfed {2.2.} and across the Welsh Basin (Robinson & Bevins 1986). During this regional event there is no evidence for temperatures having reached the amphibolite facies (>380°C), although conditions approaching 350°C are likely to have been attained, with a range in temperatures of >230 to <350°C being envisaged. The identification of greenschist assemblages west of Fishguard is at variance to the findings of others, whilst the distribution of prehnite-pumpellyite assemblages is more restricted. This may in part be reconciled by the use here of the term 'transitional' assemblage. Calculated temperatures accord well with the CAI values of 5 (≈ 300–350°C) recorded by Bergström (1980) and Savage & Basset (1985) from Abereiddi Bay. The disparity between the anchizone crystallinity recorded by Robinson *et al.* (1980) in the vicinity of Fishguard, and epizonal values recorded by Robinson & Bevins (1986) over much of SW Dyfed makes a comparison to these studies difficult. The significance of regional illite crystallinty studies is a matter of debate (see below).

A pressure estimate of probably greater than 1.2 kbars for prehnite-pumpellyite facies assemblages and possibly less than 2 kbars for greenschist facies assemblages, is at variance to the pressure estimates visualised by Bevins & Rowbotham (1983) of between 3.4 and 5 kbars for greenschist assemblages within the Welsh Basin. Their calculation is however, based on an assumed geothermal gradient of 25°C km^{-1}. This has recently be reevaluated by Robinson & Bevins (1986), using the P-T grids of Liou *et al.* (1985) and the assemblage data of Bevins & Rowbotham (1983), who suggest that conditions in the Prescelly Hills region may have reached T ≈ 315°C and P ≈ 1.85 kbars, implying a geothermal gradient of ≈ 42°C km^{-1}. Such conditions are in close agreement with the P-T conditions calculated here from assemblage development in the Fishguard district, under an envisaged geothermal gradient of 40–45°C km^{-1}. The low pressure facies series developed in the Fishguard metabasites, accords with the baric-type metamorphism envisaged by Robinson & Bevins (1986) to have been regional across the Welsh Caledonides.

Bevins & Rowbotham (1983) proposed that conditions of metamorphism in the Welsh Basin, were primarily controlled by depth of burial, related to original basin morphology and stratigraphic cover. This model is based on their recognition of a predominance of greenschist facies assemblages in metabasites from central parts of the Welsh basin, whilst marginal terrain metabasites are dominated by prehnite-pumpellyite facies assemblages. In SW Dyfed, they suggest that assemblage distribution conforms to this prograde burial model, (i.e. prehnite-pumpellyite assemblages at St. Davids/Fishguard which give way to greenschist and pumpellyite-actinolite facies assemblages in the Prescelly Hills). Robinson & Bevins (1986) substantiate the depth related model, concluding from thier regional illite crystallinity study, that isocryst values correlate directly with the original basin and shelf form.

The present author believes that sufficent evidence is now available, to suggest, firstly that generalisation of 'marginal' and 'basinal' settings are appropriate to specific areas only, and secondly, that depth related metamorphism is only part of a complex metamorphic history. For example, Bergström (1980) concluded that Mid-Ordovician CAI values of 5 and higher, are difficult to explain under the estimated stratigraphic overburden and a mean geothermal gradient. He postulated that the widespread Ordovician volcanism, or higher heat flow associated with the Caledonide Orogeny, are likely to have been conceivable additives to the background geothermal gradient. Alderidge (1986) in a study of Silurian conodont CAI values around SE Wales, draws a similar conclusion to that of Bergström (1980), calculating that an average geothermal gradient in excess of 60°C km^{-1} would be required to produce the CAI vales of 4, which he records. Alderidge (1986), expresses the opinion that Caledonian and possibly Variscan deformation may have increased the thermal flux, although he states that a tectonic thickening of 30% would be sufficient to produce the observed CAI values under a lower heat flow. The application of depth-related metamorphism is also questioned from the studies of metabasite mineralogy (Roberts 1981) and illite crystallinity (Roberts & Merriman 1985, Merriman & Roberts 1985), across the Llyn Peninsula and Snowdonia regions of North Wales. Roberts (1981) concludes, that metamorphism was syn- or immediately post-tectonic, whilst Roberts & Merriman (1985) suggests that there is a causative relationship between areas of high strain and high grade crystallinity across major folds, with isocryst patterns generally mimicking fold morphology. This indicates that tectonism (predominantly cleavage development) associated with Caledonide deformation is the major factor controlling spatial distribution and intensity of metamorphism.

It is apparent, that in studies where the conditions of metamorphism can not be adequately explained in the context of burial, the role of tectonism and deformation (Caledonide and possibly Variscan) appears to be essential in the generation of an increased heat flow. In this study evidence indicates that metamorphism is integrally related to tectonism. However, without reliable and detailed illite crystallinty data, it is suggested at present, that the tectonic influence is important is for reasons other than for instigating or elevating conditions of P–T (i.e. tectonism accentuated and enhanced phase development at the time of deformation, under P–T conditions that were pre-existing). However, if burial under a thick sedimentary pile is accountable alone for the metamorphic conditions, then it may be argued that as one progresses down through the stratigraphy an incease in metamorphic grade should become apparent. By comparing the above studies in SW Dyfed, there is no evidence to suggest this is the case.

In conclusion, whilst the P-T conditions of metamorphism in the Fishguard – Porth-gain district can be established, evidence indicates pre-, syn-, and immediate post-tectonic phase development. It is clear that metamorphism of the area is a complex interplay of factors and it is therefore difficult and possibly misleading to discuss conditions of metamorphism within this area, and the Welsh Basin, as under the influence of a single process.

CHAPTER SEVEN: SUMMARY OF FINDINGS

7.1. Summary of Findings

Presented within this thesis are several observations and interpretations regarding the Lower Palaeozoic geology between Fishguard – Porth-gain district. From each of the studied aspects, the main findings can be summarised as followed:

Styles of deformation

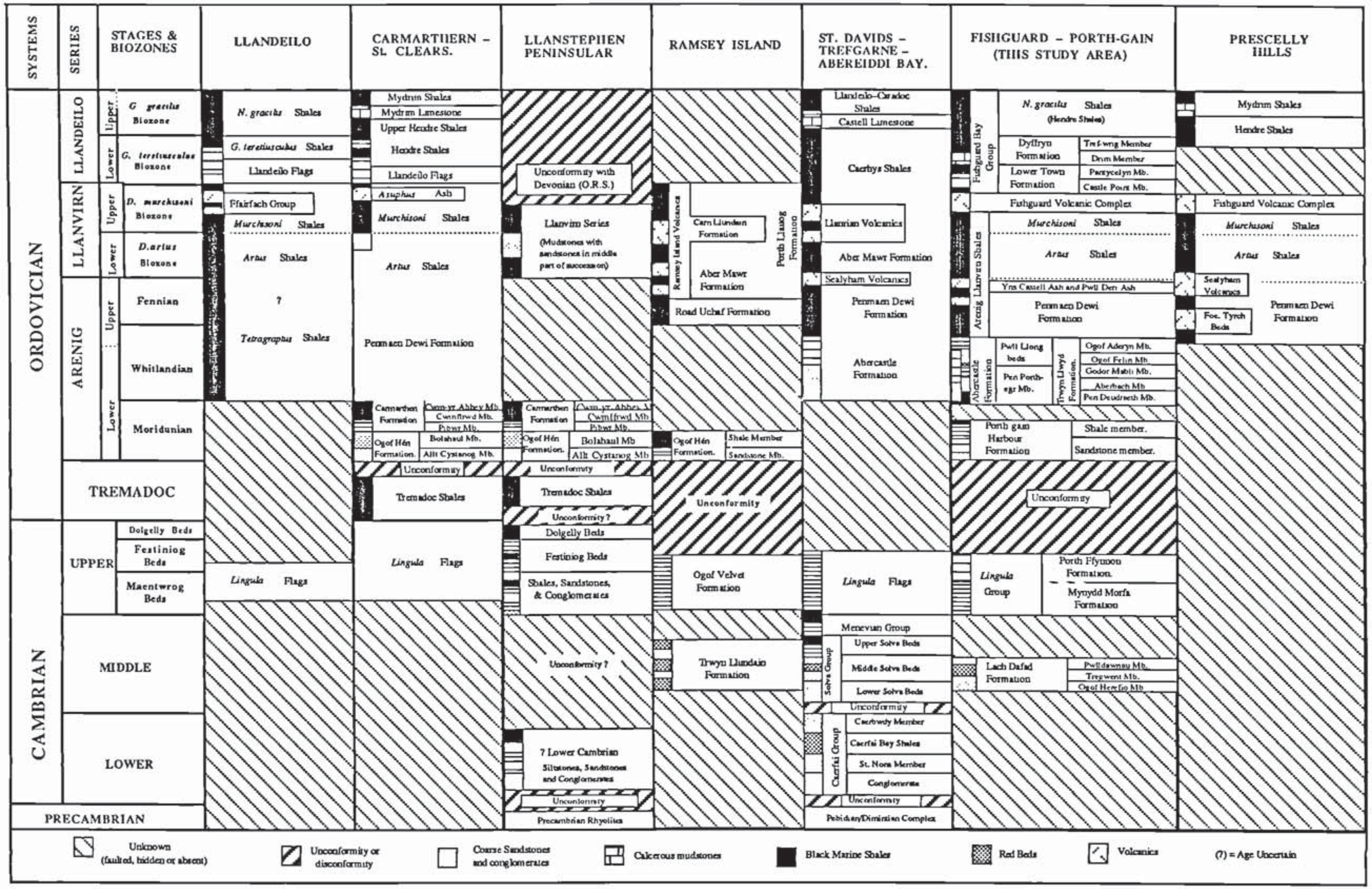
Two discrete phases of Caledonide deformation can be identified in the Fishguard – Porth-gain district. The first is seen as a regional E-W trending fold-thrust event which creates large scale repetition in the stratigraphy. On the scale of this study area, the style of deformation associated with this event is dependant on several factors, not least local anisotrophy (i.e. ductility contrasts, bedding thickness). Cleavage, folding and thrusting are approximately coeval. Thrusts commonly occur along north-facing fold limbs, reflecting accommodation in areas where folding had 'locked-up', allowing further shortening to take place. 'Stair-case' and planar bedding dislocations are also present. The direction of thrust movement is from north to south. In the regional context, it is thought that this event is not restricted to this study area. From documented evidence elsewhere, it can be speculated that deformation in the Fishguard district may represent part of a far larger regional fold-thrust belt, developing in response to shallow basement cover relationships, from Carmarthen in the east to St. Davids in the west. The second Caledonide event is expressed as NE-SW cross-faults which complicate the pre-existing structure and allochthonous stratigraphy by block faulting and rotation. The sense of movement on these latter Caledonide structures appears variable, although only locally do they have significant control on the distribution of geology. A further set of cross-faults with NW-SE trend may reflect a Variscan overprint on the Caledonide structure.

Fault rock products associated with both Caledonide events contain a wide range of low-grade metamorphic mineral assemblages. This indicates that metamorphic fluids used structural weaknesses as conduits at the time of Caledonide deformation (late-Silurian/early Devonian); therefore regional low-grade metamorphism and Caledonide deformation was in part coeval.

Sedimentary geology

The existing sedimentary stratigraphy of the Fishguard – Porth-gain district is for the most part abandoned and a new stratigraphic scheme erected (Table 3.1., 4.1.). The stratigraphy defines the regional sedimentary geology within a 'working' framework. The age of the sedimentary sequence is argued to range from the Middle Cambrian to the Lower Llandeilo, and is almost

Table 7.1. Summary of regional stratigraphic correlation.


SYSTEMS
SERIES
STAGES & BIOZONES
LLANDEILO
CARMARTHERN - St. CLEARS.
LLANSTEPHEN PENINSULAR
RAMSEY ISLAND
ST. DAVIDS - TREFGARNE - ABEREIDDI BAY.
FISHGUARD - PORTH-GAIN (THIS STUDY AREA)
PRESCELLY HILLS
ORDOVICIAN
CAMBRIAN
PRECAMBRIAN
LLANDEILO
LLANVIRN
ARENIG
TREMADOC
UPPER
MIDDLE
LOWER
Upper
Lower
G. gracilis Biozone
G. teretiusculus Biozone
D. murchisoni Biozone
D. artus Biozone
Fennian
Whitlandian
Moridunian
Dolgelly Beds
Festiniog Beds
Maentwrog Beds
N. gracilis Shales
G. teretiusculus Shales
Llandeilo Flags
Ffairfach Group
Murchisoni Shales
Artus Shales
?
Tetragraptus Shales
Lingula Flags
Mydrim Shales
Mydrim Limestone
Upper Hendre Shales
Hendre Shales
Asaphus Ash
Penmaen Dewi Formation
Carmarthen Formation
Cwm-yr-Abbey Mb.
Cwmffrwd Mb.
Pibwr Mb.
Ogof Hên Formation
Bolahaul Mb.
Allt Cystanog Mb.
Unconformity
Tremadoc Shales
Unconformity with Devonian (O.R.S.)
Llanvirn Series
(Mudstones with sandstones in middle part of succession)
Unconformity ?
Shales, Sandstones, & Conglomerates
? Lower Cambrian
Siltstones, Sandstones and Conglomerates
Precambrian Rhyolites
Ramsey Island Volcanics
Carn Llundain Formation
Aber Mawr Formation
Porth Llauog Formation
Road Uchaf Formation
Shale Member
Sandstone Mb.
Ogof Velvet Formation
Trwyn Llundain Formation
Llandeilo-Caradoc Shales
Castell Limestone
Caerbys Shales
Llanrian Volcanics
Aber Mawr Formation
Sealyham Volcanics
Penmaen Dewi Formation
Abercastle Formation
Menevian Group
Solva Group
Upper Solva Beds
Middle Solva Beds
Lower Solva Beds
Caerfai Group
Caerbwdy Member
Caerfai Bay Shales
St. Non's Member
Conglomerate
Pebidian/Dimetian Complex
N. gracilis Shales
(Hendre Shales)
Fishguard Bay Group
Dyffryn Formation
Trefwrg Member
Drim Member
Lower Town Formation
Penyrcelyn Mb.
Castle Point Mb.
Fishguard Volcanic Complex
Arenig Llanvirn Shales
Yns Castell Ash and Pwll Deri Ash
Abercastle Formation
Pwll Llong beds
Pen Porth-egr Mb.
Trwyn Llwyd Formation
Ogof Aderyn Mb.
Ogof Felin Mb.
Godor Mabli Mb.
Aberbach Mb.
Pen Deudraeth Mb.
Porth gain Harbour Formation
Shale member.
Sandstone member.
Lingula Group
Porth Ffynnon Formation
Mynydd Morfa Formation
Lach Dafad Formation
Pwlldawnau Mb.
Tregwent Mb.
Ogof Herefio Mb.
Foe. Tyrch Beds
Unknown (faulted, hidden or absent)
Unconformity or disconformity
Coarse Sandstones and conglomerates
Calcareous mudstones
Black Marine Shales
Red Beds
Volcanics
(?) = Age Uncertain

complete with regard to the development of the early Lower Palaeozoic strata of SW Dyfed. Seven principle stratigraphic units are defined and further subdivided. Regional correlation with successions elsewhere in SW Dyfed is presented, this information is summarised in Table (7.1.).

Collectively the sedimentary sequence represents an insight into ≈ 100 Ma. of sedimentation within this part of the Lower Palaeozoic Welsh Basin. Middle Cambrian sediments reflect deposition within a subtidal/tidal environment, significantly further north than may have been expected within the 'classic' Welsh Basin model. Upper Cambrian sediments reflect deposition on a storm and wave influenced, stable siliciclastic shelf which may have extended over much of South Wales. The area was uplifted during late Upper Cambrian through Tremadoc times. The onset of Ordovician sedimentation is recorded during early Arenig times by transgressive marine conglomerates and intertidal sandstones, which pass upwards rapidly into subtidal mudstones. Mid-Arenig sediments record shelf progradation and the development of a complex wave and storm dominated inner shelf depositional system within this part of SW Dyfed. Thick accumulations of turbidic mudstones reflect 'stagnation' of the marine environment during late mid-Arenig to late Upper Llanvirn times, although water depths were unlikely to have been greater than 500-1000m, and may have been appreciably less (<200m). During this period of mudstone deposition, active volcanism and high-level intrusion disrupted the argillic pile. A shallowing of the marine environment is recognised during the late Upper Llanvirn/Lower Llandeilo, possibly associated with 'pop-up' topographic highs developing in response to basement faulting. A return to marine mudstone deposition is seen within the late Lower Llandeilo and Upper Llandeilo – basal Caradoc. It is thought that the varied nature of sedimentation reflects localised change within the depositional system, syn-sedimentary tectonism (i.e. basement faulting), and the influence of larger regional eustatic events.

Igneous geology

The earliest evidence for igneous activity in the Fishguard – Porth-gain district is recorded by a volcanoclastic mass-flow (?Lahar), hosted within sediments of likely Middle Cambrian age. The onset of Ordovician igneous activity is recognised by localised accumulations of subaqueous silicic lavas, welded ash-flow tuffs, pyroclastics, and volcanoclastics of mid to late Arenig age. Hypabyssal intrusives of Arenig age are identified and may reflect the high-level counterparts of the Arenig extrusives. The episode may represent part of a larger late Arenig igneous event that was widespread throughout much of SW Dyfed.

The Fishguard Volcanic Group records Upper Llanvirn igneous activity in this part of SW Dyfed. The regional stratigraphy of the Volcanic Group has been modified (Maps 1&2). Validity for the tripartite volcanic scheme of previous workers could not be verified at outcrop, although no change to the existing nomenclature is thought warranted. One addition is the introduction of the Carreg Gybi Member, to define a series of porcellanitic mudstones, volcanoclastics and pyroclastics towards the top of the Strumble Head Volcanic Formation. The Volcanic Group records the products of both silicic and basaltic subaqueous eruptions, and high-level contemporenuous intrusives. Emplacement of lavas was below the Pressure Compensation Level for basaltic volcanism (>200m) and possibly below that for silicic lavas (?>500-1000m). The various products of volcanism reflect a complex interaction of effusive, hydroclastic, autoclastic, and gravity driven processes within the marine environment. One of the original objectives of this project was to evaluate whether or not the Fishguard Volcanic Group represents an ophiolite within the Lower Palaeozoic Welsh Basin. The present author has found no evidence to suggest that this is the case.

Metamorphic geology

Low-grade regional metamorphism of the Fishguard – Porth-gain district is identified by the development of mineral assemblages consistent with the prehnite-pumpellyite and greenschist facies. A quantitative assessment of mineral assemblage development indicates that metamorphism was of low pressure facies series with conditions in the range P = 1.2 – 2.0 kbars and T= 230 – 350°C, under an elevated geothermal gradient of between 40-45°C km^{-1}. However, metamorphism is not related to one single process, rather a range of complex processes which were pre-, syn-, and post-Caledonide deformation.

7.2. Areas for further Research

This study has investigated several aspects of the geology of the Fishguard – Porth-gain district. However, many of the areas discussed by the present author and previous workers, remain either, poorly understood, or would benefit from some further study.

A structural study of this area and surrounding districts, would do much to enhance the understanding of the Lower Palaeozoic sequence. A detailed structural study, in conjunction with biostratigraphic work, would aid any future investigation. This equally applies to large areas of SW Dyfed, which remain for the most poorly understood. The sedimentary stratigraphy has proved to be as complex as originally suggested by Cox (1930). The stratigraphy is not 'layer-cake' and marker horizons are infrequent, there being a need for the collection of undeformed diagnostic fauna to allow biostratigraphic control. This would require undoubted patience although may prove to be extremely useful both locally and regionally.

The sedimentary sequence of the district indicates a complex depositional history, which changed throughout Lower Palaeozoic times. The present author has attempted to evaluate some aspects of the sedimentology, although many successions require detailed facies analysis.

For example, entire studies could be targeted at the *Lingula* Group in this and surrounding areas of SW Dyfed. Whilst not documented in this study, brief SEM observations of sandstones from several Formations suggest a complex diagenetic/low-grade metamorphic history; studies investigating such aspects may prove to be particularly rewarding.

Igneous geology is one of the present disciplines which is in part well appreciated, previous studies have unraveled many of the problems of subaqueous volcanism, and portrayed the successions in its modern context. This is not to suggest that studies to date have been exhaustive, prolonged detailed studies may well benefit the understanding of various aspects of physical volcanology.

This study, and that of Bevins (1979), have highlighted the conditions of metamorphism by an assessment of the secondary mineral assemblages developed in basic igneous rocks. A detailed metamorphic study may be particularly rewarding, incorporating a much wider range of techniques, such as illite crystallinity, CAI, and fluid inclusion geothermometry. It would be of specific interest to evaluate the significance of illite crystallinity with regard to such sequences as the Trwyn Llwyd Formation, where areas of high- and low-strain could be compared from the same coastal sequence (e.g. Plate 3.5.a & b).

REFERENCES.

Addison, R. 1974. *The biostratigraphy of the Llandeilo facies of S. Wales*. Unpubl. Ph. D. thesis, Queens Univ. Belfast.

Aldridge, R.J. 1986. Conodont palaeobiogeography and thermal maturation in the Caledonides. *J. geol. Soc. London*. **143,** 177-184.

Allen, P.M. 1982. Lower Palaeozoic volcanism in Wales, the Welsh Borderlands, Avon, and Somerset. *In*:: **Sutherland, D.S. (ed.).** *Igneous rocks of the British Isles*. Wiley, London. 65-91.

Allen. J.R.L. 1984. Loose-boundary hydraulics and fluid mechanics: selected advances sinces 1961. *In*. **Brenchley P.J. & Williams. B.P.J.** *Sedimentology recent developments and Applied aspects*. J. geol. Soc. Spec. Publ. **18,** 7-28.

André, L., Hetrogen, J. & Deutsch, S. 1986. Ordovician-Silurian magmatic provences in Belgium and the Caledonian Orogeny in middle Europe. *Geology*. **12,** 579-582.

Andrews, D.J. & Sleep, N.M. 1974. Numerical modelling of tectonic flow behind island arcs. *Geophys. J. R. astr. Soc.* **28,** 235-251.

Anketell, J.M. 1987. On the geological successions and structure of South-Central Wales.*Geol. Jl.* **22,** 167-181.

Baker, J.W. 1982. The Precambrian of south-west Dyfed. *In* : **Basset, M.G. (ed.).** *Geological excersions in Dyfed, south-west Wales*.National Musem of Wales, Cardiff. 15-25.

Ballard, R.D. & Moore, J.G. 1977. *Photographic atlas of the Mid-Atlantic Ridge rift system valley*. Spriger-Verlag, New York.

Balatatzis, E.G. & Katagas, C.G. 1984. The pumpellyite-actinolite and contiguous facies in part of the Phyllite-Quartzite Series, central Northern Peloponnesus, Greece. *J. metamorphic Geol.* **2,** 349-363.

Banks, N.L. 1973. Innerelv Member: Late Precambrian marine shelf deposit, East Finnmark. *Norg. geol. Unders.* **288,** 7-25.

Baragar, W.R.A., Plant, A.G., Pringle, G.J. & Schau, M. 1979. Diagenetic and post diagenetic changes in the comparison of an Archean Pillow. *Can. J. Earth Sci.* **16,** 2102-2121.

Basset, M.G. (ed.). 1982. *Geological excersions in Dyfed, south-west Wales*. National Museum of Wales, Cardiff. 327 pp.

Bates, D.E. 1969. Some early Arenig brachiopod and trilobite faunas from Wales. *Bull. Br. Mus. nat. Hist. Ser. Geol.* **18,** 193-208.

Beckly, A.J. 1987. Basin development in North Wales during the Arenig. *Geol. Jl.* **22,** 19-30.

Bergström, S.M. 1971. Conodont biostratigraphy of the Middle and Upper Ordovician of Europe and North America. *Mem. geol. Soc. Am.* **127,** 83-157.

Bergström, S.M. 1973. Correlation of the late Lasnasnamagian Stage (Middle Ordovician) with the graptolite succession. *Geol. Foren. i Stockholm Forh.* **95,** 9-18.

Bergström, S.M. 1980. Conodonts as palaeotemperature tools in the Ordovician of the Caledonides and adjacent areas in Scandanavia and the British Isles. *Geol. Foren. i Stockholm Forh.* **102,** 377-392.

Bevins, R.E. 1978. Pumpellyite-bearing igneous rocks from the Lower Ordovician of North Pembrokeshire, Wales. *Min. Mag. Lond.* **42,** 81-83.

Bevins, R.E. 1979. *The geology of the Strumble Head - Fishguard region, Dyfed, Wales.* Unpubl. PhD thesis. Univ. Keele.

Bevins, R.E. 1982. Petrology and geochemistry of the Fishguard Volcanic Complex, Wales. *Geol. J.* **17,** 1-21.

Bevins, R.E. 1985. Pumpellyite-dominated metadomain alteration at Builth Wells, Wales; evidence for a fossil submarine hydrothermal system ?. *Min. Mag. Lond.* **49,** 451-456.

Bevins, R.E., Kokelaar, B.P. & Dunkley P.N. 1984. Petrology and geochemistry of lower to middle Ordovician igneous rocks in Wales: a volcanic arc to marginal basin transition. *Proc. geol. Assoc. London.* **95,** 337-347.

Bevins, R.E. & Roach, R.A. 1979a. Early Ordovician volcanism in Dyfed, S.W. Wales. *In:* **Harris, A.L., Holland, C.H. & Leake, B.E. (eds.).** *The Caledonides of the British Isles - reveiwed.* Spec. Publ. geol. Soc.London, 8, 603-609.

Bevins, R.E. & Roach, R.A. 1979b. Pillow lavas and isolated-pillow breccia of rhyodacite composition from the Fishguard Volcanic Group, Lower Ordovician, S.W. Wales, U.K. *J. Geol.* **87,** 193-201.

Bevins, R.E. & Roach, R.A. 1982. Ordovician igneous actvity in south-west Wales. *In:* **Basset, M. (ed).** *Geololgical excursions in Dyfed, south-west Wales.* National Museum of Wales, Cardiff. 65-80.

Bevins, R.E. & Rowbotham, G. 1983. Low grade regional metamorphism within the Welsh sector of the paratectonic Caledonides. *Geol. J.* **18,** 141-167.

Bird, D.K. & Helgeson H.C. 1981. Chemical interractions of aqueous solutions with epidote-feldspar mineral assemblages in geological systems. Equilibrium constaints in metamorphic/ geothermal processes. *Am. J. Sci.* **281,** 576-614.

Black, W.W., Bulman, O.M.B., Hey, R.W. & Hughes, C.P. 1971. Ordovician stratigraphy of Abereiddy Bay, Pembrokeshire.*Geol. Mag.* **10,** 546-548.

Boles, J.R. & Coombs, D.S. 1977. Zeolite facies alteration of sandstones in the Southland Syncline, New Zealand. *Am. J. Sci.* **277,** 982-1012.

Bowen, D.Q. 1982. Pleistocene deposits and fluvio-glacial landforms of North Preseli. *In* : **Basset, M.G. (ed.).** *Geological excersions in Dyfed, south-west,Wales.* National Musem of Wales, Cardiff. 289-295.

Brenchly, P.J. & Newall, G. 1982. Strom-influenced inner-shelf sand-lobes in the Caradoc (Ordovician) of Shropshire, England. *Jl. Sed. Petrol.* **52,** 1257-1269.

Brown, E.H. 1967. The greenschist facies in part of eastern Otago, New Zealand.*Contrib. mineral. Petrol.* **14,** 259-292.

Brown, E.H. 1971. Phase relations of biotite and stilpnomelane in the greenschist facies. *Contrib. mineral. Petrol.* **31,** 275-299.

Brown, E.H. 1977. Phase equilibria amoung pumpellyite, lawsonite, epidote and associated minerals in low-grade metamorphic rocks. *Contrib. mineral. Petrol.* **46,** 123-136.

Brown, P.E., Tocher, F.E. & Chambers, A.D. 1982. Amphiboles in the Lilloise intrusion, East Greenland. *Min. Mag. London.* **45,** 47-54.

Bryan, W.B. 1972. Morphology of quench crystals in submarine basalts. *J. geophys. Res.* **77,** 5812-5819.

Butler, R.W.H. 1982. The terminology of structures in thrust sheets. *Jl. strucural. Geol.* **4,** 239-245.

Cann, J.R. 1979. Metamorphism in the ocean crust.*In*: **Talwani, M. Harrison, C.G. & Hayes, D.R.** (eds.). *DSDP drilling results in the Atlantic Ocean, Maurice Ewing Series 2.* American Geophysical Union, Scientific Report No 48, 230-238.

Carey, S. & Sigurdsson, H. 1984. A model of volcanogenic sedimentation in marginal basins. *In* : **Kokelaar, B.P. & Howells, M.F.** (eds.) *Marginal Basin Geology, Volcanic and Associated Sedimentary and tectomic Processes in Modern and Ancient Marginal Basins.* Spec. Publ. geol. Soc. Lond. **16,** 37-58.

Carlisle, D. 1963. Pillow breccias and their aquagene tuffs, Quandra Island, British Columbia. *J. Geol.* **71,** 48-71.

Carter, C.W. 1978. A regressive barrier and barrier-protected deposit: depositional environments and geographic setting of the late Tertiary Cohansey Sand. *Jl. Sed. Petrol.* **48,** 933-950.

Cas, R. 1978. Silicic lavas in Palaeozoic flyschlike deposits in New South Wales, Australia: Behaviour of deep subaqueous silicic flows. *Geol. Soc. Am. Bull..* **89,** 1709-1714.

Chafetz, H.S. 1982. The Upper Cretaceous Beartooth Sandstone of Southwestern New Mexico: A transgressive deltaic complex on silicified palaoeskarst. *Jl. Sed. Petrol.* **52,** 157-169

Cho, M., Liou, J.G. & Maruyama, S. 1986. Transition from the zeolite to the prehnite-pumpellyite facies in the Karmutsen Metabasites, Vancouver Island, British Columbia. *J. Petrol.* **27,** 467-494.

Cole, J.W. 1985. Taupo-Rotorua Depression: an ensialic marginal basin of North Island, New Zealand. *In*: **Kokelaar, B.P. & Howells, M.F.** (eds.). *Marginal Basin Geology: Volcanic and Associated Sedimentary and Tectonic Processes in Modern and Ancient Marginal Basins.* Spec. publ. geol. Soc. London. **16,** 109-120

Coleman, J.M. & Prior, D.B. 1980. *Deltaic Sand Bodies.* Course note series 15, Bull. Am. Ass. Petrol. geol., Tulsa, Oklahoma. p.171.

Coombs, D.S. 1960. Low grade mineral facies in New Zealand. *Rep. international. Geol. Congr. 21st Session.* **13,** 339-351.

Coombs, D.S., Kawachi, Y., Houghton, B.F., Hyden, G., Pringle, I.J. & Williams, J.G. 1977. Andradite and andradite-grossular solid solution in very low grade regionally metamorphosed rocks in southern New Zealand. *Contrib. mineral. Petrol.* **63,** 229-246.

Coombs, D.S., Nakamura, Y. & Vuagnat, M. 1976. Pumpellyite-actinolite facies schists of the Taveyanne Formation near Loeche, Valais, Switzerland. *J. Petrol.* **17,** 440-471.

Cooper, R.A. & Fortey, R.A. 1982. The Ordovician graptolites of Spitsbergen. *Bull. Br. mus. Nat. Hist. Lond. Geol Ser.* **36,** 157-302.

Cope, J.W.C. 1979. Early history of the southern margin of the Tywi anticline in the Carmarthen area, South Wales. *In* : **Harris, A.L., Holland, C.H. & Leake, B.E.** (eds.). *Caledonides of the British Isles - reviewed.* Scottish Academic Press. 527-532.

Cope, J.W.C., Fortey, R.A. & Owens, R.M. 1978. Newly discovered Tremadoc rocks in the Carmarthern district, South Wales. *Geol. Mag.* **115,** 195-198.

Cortesogno, L. & Lucchetti, G. 1984. Ocean floor metamorphism of metagabbro's and striped amphibolites, (T.Murlo Southern Tuscany, Italy). *Neues Jahrbuch. Miner. Abh.* **148,** 276-300.

Cotter, E. 1985. Gravel-topped offshore bar sequences in the Lower Carboniferous of southern Ireland. *Sedimentology.* **32,** 195-213.

Cowie, J.W., Ruston, A.W.A. & Stubblefield, C.J. 1972. A correlation of Cambrian rocks in the British Isles. *Geol. Soc. London. Spec. Rep.* **2,** 1-42.

Cowie, J.W. & Rushton, A.W.A. 1974. Palaeogeography of the Cambrian in the British Isles *In:* **Holland, C.H.** (ed.). *The Cambrian of the British Isles, Norden and Spitsbergen (Lower Paleozoic rocks of the world, 2).* Wiley & Sons. 177-183.

Cox, A.H. 1915. The geology of the district between Abereiddy Bay and Abercastle. *Q. J. geol. Soc. London.* **71,** 273-340.

Cox, A.H. 1930. A preliminary note on the geological structure of Pen Caer and Strumble Head, Pembrokeshire. *Proc. geol. Assoc. London.* **41,** 274-289.

Cox, A.H. & Jones, O.T. 1913. "Pillow-lavas in North and South Wales". *Rep. Brit. Assoc. Birmingham;* p 495.

Cox, A.H., Green, J.F.N., Jones, O T., & Pringle, J. 1930a. The geology of the St. Davids district, Pembrokeshire. *Proc. geol. Assoc. London.* **41,** 241-273.

Cox, A.II., Green, J.F.N., Jones, O.T. & Pringle, J. 1930b. The St. Davids district: report of the summer meeting 1930. *Proc. geol. Assoc. London.* **41,** 412-438.

Crimes, T.P. 1970. A facies analysis of the Cambrian of Wales. *Palaeogeog. Palaeoclim. Palaeoecol.* **7,** 113-152.

Dahlstrom, C.D.A. 1970. Strucural geology in the eastern margin of the Canadian Rocky Mountains. *Bull. Can. Petrol. Geol.* **18,** 332-406.

Davies, D.K., Ethridge, F.G. & Berg, R.R. 1971. Recognition of barrier environments. *Bull. Am. Ass. Petrol. geol.* **55,** 550-565.

Deer, W.A., Howie, R.A. & Zussman, J. 1962. *Rock Forming Silicates (5volumes).* Longmans Group Ltd, London.

Deer, W.A., Howie, R.A. & Zussman, J. 1982. *Rock-Forming Minerals, Orthosilicates Volume 1A.* Longman Group Ltd, London.

Dimroth, E., Cousineau, P., Luduc, M. & Sanschagrin, Y. 1978. Structure and organization of Archean subaqueous basaltic flows, Rouyn-Noranda area, Quebec, Canada. *Can. J. Earth Sci.* **15,** 902-918.

Dorjes, J. & Hertweck, G. 1975. Recent biocoenoses and ichnocoenoses in shallow-water environments. *In:* **Frey, R.W. (ed).** The study of Trace Fossils, 459-491. Springer-Verlag, New York.

Duffield, W.A. 1969. Concentric structures in elongated pillows, Amador County, California. *U.S. geol. Surv. Prof. Pap.* **650-D,** 19-25.

Duke, W.L. 1985. Hummocky cross-stratification, tropical hurracanes, and intense winter storms. *Sedimentology.* **32,** 167-194.

Dunne, W.M. 1983. Tectonic evolution of SW Wales during the Upper Palaeozoic. *J. geol. Soc. London.* **140,** 257-265.

Eggleton, R.A. 1972. The crystal structure of stilpnomelane. (II). The subcell. *Min. Mag. Lond.* **38,** 693-711.

Elliot, T. 1978a. Deltas. *In:* **Reading H.G. (ed).** *Sedimentary Environments and Facies.* Blackwell Scientific Publications. 97-142.

Elliot, T. 1978b. Clastic Shorelines. *In:* **Reading H.G. (ed).** *Sedimentary Environments and Facies.* Blackwell Scientific Publications. 143-177.

Elsden, J.V. 1905. On the igneous rocks ocurring between St. Davids Head and Strumble Head. *Q. J. geol. Soc. Lond.* **61,** 579-607.

Epstein, A.G., Epstein, J.B. & Harris, L.D. 1977. Conodont color alteration - as an index to organic metamorphism. *U.S. geol. Surv. Prof. Paper.* **995,** 27 pp.

Either, V.G. & Cambell, R.G. 1977. Tourmaline concetrations in Protorozoic sediments of the southern Cordillera of Canada and their economic significance. *Can. J. Earth Sci.* **14,** 2348-2363.

Evans, W.D. 1945. The geology of the Prescelly Hills, North Pembroke- shire. *Q. J. geol. Soc. Lond.* **101,** 89-107.

Evarts, R.C. & Schiffman, P. 1983. Submarine hydrothermal metamorphism of the Del Puetro Ophiolite, California. *Am. J. Sci.* **283,** 289-340.

Feininger, T. 1984. Stilpnomelane in metasomatic rocks associated with steatite in regional schists, Quebec Appalachians. *Can. Mineral.* **22,** 423-435.

Fisher, R.V. 1983. Flow transformation in sedimentary gravity flows. *Geology.* **11,** 273-274.

Fisher, R.V. 1984. Submarine volcanoclastic rocks. *In:* **Kokelaar, B.P. & Howells, M.F. (eds.).** *Marginal Basin Geology; Volcanism and associated sedimentary processes in modrn and ancient marginal basins.* Spec. Publ. geol. Soc. London. **16,** 5-28.

Fisher, R.V. & Schmincke, H-U. 1984. *Pyroclastic Rocks.* Springer-Verlag, Heidelberg. pp 472.

Fiske, R.S. 1963. Subaqueous pyroclastic flows in the Ohanapecosh Formation, Washington. *Bull. geol. Soc. Am.* **74,** 391-406.

Fiske, R.S. & Matsuda, T. 1964. Submarine equivalents of ash flow tuffs on the Tokiwa Formation, Japan. *Am. J. Sci.* **262,** 76-106.

Fitches, W.R. 1972. Polyphase deformation structures in the Welsh Caledonides near Aberystwyth.*Geol. Mag.* **109,** 149-155.

Fitches, W.R., Cave, R., Craig, J. & Maltham, A.J. 1986. Early veins as evidence of detachment in the Lower Palaeozoic rocks of Wales. *J. structural. Geol.* **8,** 607-620.

Fitches, W.R. & Cambell, S.D.G. 1987. Tectonic evolution of the Bala Fault Lineament in the Welsh Basin. *Geol. Jl.* **22,** 131-153.

Fitches, W.R. & Woodcock, N.H. 1987. *Conference report;* Sedimentation and tectonics in the Welsh Basin. *J. geol. Soc. Lond.* **144,** 199-200.

Foranari, D.J., Malahoff, A. & Heezen, B.C. 1979. Visible observations of the volcanic micromorphology of Tortuga, Lorraine and Tutu seamounts: petrology/geochemistry of ridge and seamount features in and around the Panama Basin. *Mar. Geol.* **31,** 1-30

Fortey, R.A. & Cocks, L.R.M. 1986. Marginal faunal belts and their structural implications. *J. geol. Soc. London.* **143,** 151-160.

Fortey, R.A. & Owens, R.M. 1978. Early Ordovician (Arenig) stratigraphy and faunas of the Carmarthen district, south-west Wales. *Bull. Br. Mus. nat. Hist. Ser. Geol.* **30,** 225-294.

Fortey, R.A. & Owens, R.M. 1987. The Arenig Series in South Wales. *Bull. Br. Mus. nat. Hist. Ser. Geol.*

Francis, E.H. 1983. Magma and sediment – II. Problems in interpreting palaeovolcanics in the stratigraphic column. *Jl. geol. Soc. Lond.* **140** 165-183.

Furnes, H. 1973. Variolitic stucture in Ordovician pillow lava and its possible significance as an environmental indicator. *Geology.* **1,** 27-30.

Furnes, H. & Frideifsson, I.B. 1979. Pillow block breccia - occurrance and mode of formation. *Neues. Harbuch. fur. Geologhe. und. Palentologie. Montashefte.* **3,** 147-154.

Fyfe, W.S., Price, N.J. & Thompson, A.B. 1978. *Fluids in the earths crust.* Amsterdam, Elsevier. 383pp.

George, T.N. 1970. *British Regional Geology: South Wales*. 3rd ed. Inst. Geol. Sci. H.M.S.O. London. 152pp.

Glassey, W.E. 1975. Low varience phase relationships in a prehnite-pumpellyite facies terrain. *Lithos*. **8,** 69-76.

Goldering, R. & Bridges, P. 1973. Sublittoral sheet sandstones. *Jl. Sed. Petrol.* **43,** 736-747.

Gole, M.J. 1980. Low-temperature retrograde minerals in metamorphosed Archean banded iron-formations, western Australia. *Can. mineral.* **18,** 205-214.

Graham, C.M. 1976. A note on some Dalradian stilpnomelanes and their oxidation. *Min. Mag. Lond.* **40,** 467-472.

Grapes, R.H. & Graham, C.M. 1976. The actinolite-hornblende series in metabasites and the so-called miscibility gap: a reveiw. *Lithos*. **11,** 85-97.

Green, J.N.F. 1908. The geological stucture of the St. Davids area (Pembrokeshire). *Q. J. geol. Soc. Lond.* **64,** 363-368.

Greene, T. & Roberts, D. 1983. Volcanostratigrahy and eruptive products of the Jonsvaten Greenstone Formation, central Norwegian Caledonides. *Nor. geol. unders.* **387,** 21-38.

Hall, R.P. 1985. Mg-Fe-Mn distribution in amphiboles, pyroxenes and garnets and implications for conditions of metamorphism of high grade early Archean iron- formations, south-west Greenland. *Min. Mag. Lond.* **49,** 117-128.

Hanson, R.E. & Schweickert, R.A. 1982. Chilling and brecciation of a Devonian rhyolite sill intruded into wet sediments, northrn Sierra Nevada, California. *J. Geol.* **90,** 717-724.

Hashimoto, M. 1972. Reactions producing actinolite in basic metamorphic rocks. *Lithos*. **5,** 19-31.

Hawthorn, F.C. 1981. Crystal chemistry of the amphiboles. In: *Amphiboles and other hydrous pyriboles - reviews in mineralogy*. Min. Soc. Am. Spec. Pub. **9A,** 1-95.

Hey, M.M. 1954. A new review of the chlorites. *Min. Mag. Lond.* **30,** 277-292.

Hicks, H. 1873. On the Tremadoc rocks in the neighbourhood of St. Davids. *Q. J. geol. Soc. Lond.* **29,** 39-52.

Hicks, H. 1875. On the succession of the ancient rocks in the vicinity of St. Davids, Pembrokeshire, with special reference to those of the Arenig and Llandeilo Groups, and their fossil contents. *Q. J. geol. Soc. Lond.* **31,** 167-195

Hicks, H. 1881. The classification of the Eozoic and Lower Palaeozoic rocks of the British Isles. *Pop. Sci. Review*. **5,** 289-308.

Higgins, J.B. & Rubbie, P.M. 1976. The crystal stucture and space groups of natural and synthetic titanites. *Am. Mineral.* **61,** 878-888.

Holland. C.H. ***et al.*** ***(+17).*** A guide to stratigraphic procedure. *Geol. Soc. Lond. Spec. Rep.* **10.**

Hoffman, H.J. 1975. *Bolopora* not a bryozoan, but an Ordovician phosphatic, oncoltic accretion. *Geol. Mag.* **112,** 523-526

Hoffman, J. & Hower, J. 1979. Clay mineral assemblages as low-grade metamorphic geothermometers; Application to the thrust faulted disturbed belt of Montana, U.S.A. *In:* **Scholle, P.A. & Schluger, P.R. (eds.).** *Aspects of Diagenesis.* Soc. Econ. Palaeo. Mineral. Spec. Publ. **26,** 55-79.

Honnorez, J. & Kirst, P. 1975. Submarine basaltic volcanism: morphometric parameters for discriminating hyaloclastites from hyalotuffs. *Bull. Volcanol.* **39,** 441-65.

Houghton, B.F. 1982. Low-grade metamorphism of the Takitimu Group, western Southland, New Zealand. *N. Z. J. geol. Geophys.* **25,** 1-19.

House, M.R., Richardson, J.B., Chaloner, W.G., Allen,. J.R.L., Holland, C.H. & Westoll, T.S. 1977. A correlation of Devionian rocks of the British Isles. *Geol. Soc. Lond. Spec. Report.* **8,** 110 pp.

Howard, J.D. 1971. Comparison of the beach and to shoreface sequence in modern and ancient sediments. In: *Recent Advances in Paleocology and Icnology.* Amer. Geol. Inst. 148-183.

Howard, J.D. & Reineck, H-E. 1981. Depositional facies of high-energy beach to offshore sequence: Comparison with low-energy sequence. *Bull. Am. Ass. Petrol. geol..* **65,** 807-830.

Howells, M.F., Cambell, D.G. & Reedman, A.J. 1985. Isolated pods of subaqueous welded ash-flow: a distal facies of the Capel Currig Volcanic Formation (Ordovician), North Wales. *Geol. Mag.* **122,** 175-180.

Hughes, C.P., Jenkins, C.J. & Rickards, R.B. 1982. Abereiddi Bay and the adjacent coast. *In*:: **Basset, M.G. (ed.).** *Geological excersions in Dyfed, south-west Wales.* National Museum of Wales. 51-63.

Humphris, S.E. & Thompson, G. 1978. Hydrothermal alteration of oceanic basalts by seawater. *Geochem. cosmochem. Acta.* **42,** 107-125.

Hynes, A. 1982. A comparison of amphiboles from medium and low-grade metabasites. *Contrib. mineral. Petrol.* **81,** 119-125.

Johnson, H. D. 1978. Shallow Siliciclastic Scas. *In:* **Reading H.G. (ed).** *Sedimentary Environments and Facies.* Blackwell Scientific Publications. 207-258.

Johnson, H. D. & Stewart D.J. 1984. Role of clastic sedimentology in the exploration and production of oil and gas *In.* **Brenchley P.J. & Williams. B.P.J.** *Sedimentology recent developments and Applied aspects.* J. geol. Soc. Spec. Publ. **18,** 7-28.

Jones, J. G. 1969. Pillow lavas as depth indicators. *Am. J. Sci.* **251,** 181-195.

Jones, O. T. 1940. Some Lower Palaeozoic contacts in Pembrokeshire. *Geol. Mag.* **77,** 405-409.

Kastner, M. & Siever, R. 1983. Siliceous sediments of the Guaymas Basin: The effect of high thermal gradients on diagenesis. *J. Geol.* **91,** 629-641.

Katagas, C. & Panagos, A.G. 1979. Pumpellyite-Actinolite and greenschist facies metamorphism in Levos Island (Greece). *Tschermaks. mineral. petrol. Mitt.* **26,** 235-254.

Kawachi, Y. 1975. Pumpellyite-actinolite and contiguous facies metamorphism in parts of the Upper Wakatipu district, Southern Island, New zealand. *N.Z.J. geol. Geophys.* **18,** 401-441.

Kay, S. M. 1983. Metamorphism in the Alutian arc: the Finger Bay Pluton, Adak, Alaska. *Can. Mineral.* **21,** 665-681.

Kennedy, M.J. 1979. The continuation of the Canadian Appalacians into the Caledonides of Britain and Ireland. *In*: **Harris, A.L., Holland, C.H. & Leake, B.E. (eds.).** *The Caledonides of the British Isles - reviewed.* Spec. Publ. geol. Soc. Lond. **8,** 33-64.

Kennedy, R.J. 1986. *Llanvirn trilobites of South Wales.* Unpubl. M.Phil thesis. Univ. Aston. Birmingham.

Kennedy, R.J. (in press). Llanvirn trilobites of S Wales. *Palaeoentology.*

Klein, G. DE V., 1971. A sedimentary model for determining palaeotidal range. *Geol. Soc. Am. Bull.* **82,** 2585-2592.

Klien, C.Jr. 1974. Greenalite, stilpnomelane, minnesotaite, crocodilite and carbonates in a very low grade metamorphic Precambrian iron-formation. *Can Mineral.* **12,** 475-498.

Kirkpatrick, R.J. 1979.a. Petrology of basalts: Hole 369B, DSDP Leg 46. *In*: **Dmitriev, L. *et al.* (eds).** Inital Reports of the Deep Sea Drilling Project. **46,** Washington D.C., U.S. Government Printing Office p165 - 178.

Kokelaar, B.P. 1982. Fluidization of wet sediments during the emplacement and cooling of various igneous bodies. *J. geol. Soc. Lond.* **139,** 21-33.

Kokelaar, B.P., Howells, M.F., Bevins, R.E., Roach, R.A. & Dunkley, P.N. 1984.a. The Ordovician marginal basin of Wales; *In*: **Kokelaar, B.P. & Howells, M.F. (eds.)** *Marginal Basin Geology, Volcanic and Associated Sedimentary and Tectonic Processes in Modern and Ancient Marginal Basins.* Spec. Publ. geol. Soc. Lond. **16,** 245 - 269.

Kokelaar, B.P., Howells, M.F., Bevins, R.E. & Roach, R.A. 1984.b. Volcanic and associated sedimentary and tectonic processes in the Ordovician Marginal Basin of Wales, a field guide. *In*:: **Kokelaar, B.P. & Howells, M.F. (eds).** *Marginal Basin Geology,Volcanic and Associated Sedimentary and Tectonic Processes in Modern and Ancient Marginal Basins.* Spec. Publ. geol. Soc. Lond. **16,** 291-322.

Kokelaar, B.P., Bevins, R.E. & Roach, R.A. 1985. Submarine silicic volcanism and associated sedimentary and tectonic processes, Ramsey Island, SW Wales. *J. geol. Soc. Lond.* **142,** 591 - 613.

Korikovskiv, S.P., Boronikhin, V.A. & Laputina, I.P. 1975. High temperature boundary of the stability field of stilpnomelane in the kyanite metamorphic complex Potam Mountains. *Acad. Sci. USSR Dokl., Earth Sci.* Sect **222,** 205-207.

Kuniyoshi, S. & Liou, J.G. 1976. Contact metamorphism of the Karmutsen Volcanics, Vancouver Island, British Columbia. *J. Petrol.* **17,** 73-99.

Laird, J., Lanphere, M. & Albee, A. 1984. Distribution of Ordovicain and Devonain metamorphism in mafic and pelitic schists from northern Vermont. *Am. J. Sci.* **281,** 127-175.

Lake, P. 1906. British Cambrian Trilobites. *Mono palaeontogr. Soc. London.* **(1),** 1-28.

Lake, P. 1908. British Cambrian Trilobites. *Mono palaeontogr. Soc. London.* **(3),** 49-64.

Leake, B.E. 1965. The relationship between tetrahedral aluminium and the maximum possible octahedral aluminium in natural calciferous and subcalciferous amphiboles. *Am. Mineral.* **50,** 843-851.

Leake, B.E. 1978. Nomenclature of amphiboles. *Min. Mag. Lond.* **42,** 533-563.

Leeder, M.R. 1982. *Sedimentology: Process and products.* George, Allen & Unwin Ltd.

Legget, J.K. 1980. British Lower Palaeozoic black shales and their palaeo-oceanographic significance. *J. geol. Soc. London.* **137,** 139-156.

Liou, J.G. 1971. Synthesis and stability relations of prehnite $CaAl_2Si_3O_{10}$ $(OH)_2$. *Am. Mineral.* **56,** 507-531.

Liou, J.G. 1979. Zeolite facies metamorphism of basaltic rocks from the east Taiwan Ophiolite. *Am. Mineral.* **68,** 335-348.

Liou, J.G. & Ernst, W. G. 1979. Oceanic ridge metamorphism of the east Taiwan Ophiolite. *Contrib. mineral. Petrol.* **68,** 335-348.

Liou, J.G., Kim, H.S. & Maruyama, S. 1983. Prehnite-epidote equilibria and their petrologic applications. *J. Petrol.* **24,** 321-342.

Liou, J.G., Maruyama, S. & Cho, M. 1985. Phase equilibria and mineral paragenesis of metabasites in low-grade metamorphism. *Min. Mag. Lond.* **49,** 321-33.

Lonsdale, P. & Baltiza, R. 1980. Hyaloclastite and lava flows on young seamounts examined with a submersible. *Geol. Soc. Am. Bull.* **91,** 545-554.

Lopez-Montano, R. 1983. Stilpnomelane and associated minerals in the Hällefors dolerite dyke, south central Sweden. *Geologiska Fören .i. Stock Förhand.* **106,** 73-79.

Lorenz, B.E. 1984. Mud-magma interaction in the Dunnage Mélange, Newfoundland. *In:* **Kokelaar, B.P. & Howells, M.F. (eds.).** *Marginal Basin Geology, Volcanic and Associated Sedimentary and Tectonic Processes in Modern and Ancient Marginal Basins.* Spec. Publ. Geol. Soc. London. **16,** 271-277.

Lowe, D.R. 1979. Sedimentary gravity flows: their classification and some problems of application to natural flows and deposist. *Spec. Publ. Soc. econ. Paleontol. Mineral., Tulsa.* **27,** 75-82.

Lowe, D.R. 1982. Sediment gravity flows. II. Depostional models with specific reference to deposits of high-density turbidity currents. *Jl. sed. Petrol.* **52,** 279-297.

Lowman, R.D.W. 1977. *The geology, petrology, and geochemistry of an area of Lower Palaeozoic rocks, east of Fishguard, North Pembrokeshire.* Unpubl. PhD Thesis, Univ of Wales.

Lowman, R.W.D. & Bloxam, R.W. 1981. The petrology of the Lower Palaeozoic Fishguard Volcanic Group and associated rocks, E. of Fishguard, N. Pembrokeshire, Dyfed, South Wales. *J. geol. Soc. London.* **129,** 481-503.

Maltman, A. 1984. On the term 'soft-sediment deformation'.*Jl. structural Geol.* **6,** 589-592.

Maruyama, S. Suzuki, K. & Liou, J.G. 1983. Greenschist-Amphibolite transition equilibria at low pressures. *J. Petrol.* **24,** 583-604.

McKerrow, W.S. & Cocks, L.R.M. 1986. Oceans, island arcs and olistostromes: the use of distinguishing sutures, terranes and environments around the Iapetus Ocean. *J. geol. Soc. Lond.* **143,** 185-191.

Merriman, R.J. & Roberts, B. 1985. A survey of white-mica crystallinity and polytypes in pelitic rocks of Snowdonia and Llyn, North Wales. *Min. Mag. Lond.* **49,**305-319.

Mevel, C. 1981. Occurance of pumpellyite in hydrothermally altered basalts from the Verma Fracture Zone (Mid-Atlantic Ridge). *Contrib. mineral. Petrol.* **76,** 386-393.

Middleton, G.V. 1967. Experiments on density and turbidity currents (III). Deposition of sediment. *Can. J. Earth Sci.* **4,** 475-505.

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 & Sons 197-218.

Mills, A.A. 1984. Pillow lavas and the Liedenfrost effect. *J. geol. Soc. Lond.* **141,** 183-186.

Moore, D.E. 1984. Metamorphic history of a high-grade blueschist exotic block from the Franciscan Complex, California. *J. Petrol.* **25,** 126-150.

Moore, J.G. 1975. Mechanisms of the formation of pillows. *Am.Scientist.* **63,** 269-277.

Mount, J.F. 1982. Storm-surge-ebb origin of hummocky cross-stratification units of the Andrews Mountain Member, Camptio Formation (Lower Cambrian), White-inyo Mountains Easten California. *Jl.Sed. Petrol.* **52,** 941-958.

Munhá, J., Fyfe, W.S. & Kerrich, R. 1980. Adularia, the characteristic mineral of felsic spilites. *Contrib. mineral. Petrol.* **75,** 15-19.

Naylor, M. A. 1980. The origin of inverse grading in muddy debris flow deposits – a review. *J. sed. Petrol.* **50,** 1111-1116.

Niem, A.R. 1977. Mississippian pyroclastic flow and ash-fall deposits in the deep-marine Ouachita flysch basin, Oklahoma and Arkansas. *Bull. geol. Soc. Am.* **88,** 49-61.

Nitsch, K.H 1971. Stabilitätsbeziehungen von prehnit und pumpellyitehaltigen paragenses. *Contrib. mineral. Petrol.* **30,** 240-260.

Nyström, J.O. 1983. Pumpellyite-bearing rocks in central Sweden and extent of host rock alteration as a control of pumpellyite composition.*Contrib. mineral. Petrol.* **83,** 159-168.

Offler, R. 1984. Subcalcic, Fe-rich amphiboles in metadolerites, Glenrock Station, NSW, Australia. *Min. Mag. Lond.* **48,** 47-52.

Offler, R., Baker, C.K. & Gamble, J. 1981. Pumpellyites in two low grade metamorphic terranes north of Newcastle, NSW, Australia. *Contrib. mineral. Petrol.* **76,** 171-176.

Oliver, G.J.H. & Legget, J.K. 1980. Metamorphism in an accretionary prism, prehnite-pumpellyite facies metamorphism of the Southern Uplands of Scotland. *Trans. Roy. Soc. Edinb: Earth Sciences.* **71,** 235-246.

Oomkens. E. 1967. Depositional sequences and sand bar distribution in a deltaic complex. *Geol. Minjnb.* **46,** 265-278.

Part, G.M. 1922. Notes on the Ordovician lavas of Mynydd Prescelly, N. Pembrokeshire *Geol. Mag.* **59,** 310-323.

Passaglia, E. & Gottardi, G. 1973. Crystal chemistry and nomenclature of pumpellyite and julgoldites. *Can. Mineral.* **12,** 219-223.

Patchett, P.J. & Jocelyn, J. 1979. U–Pb zircon ages for late Precambrian igneous rocks in South Wales. *J. geol. Soc. Lond.* **136,** 13-19.

Pedersen, G.K. 1985. Thin, fine-grained storm layers in a muddy shelf sequence: an example from the Lower Jurrassic in the Stenlille 1 well, Denmark. *Jl. geol. Soc. Lond.* **142,** 357-374.

Pickering, K.T. 1982. A Precambrian upper basin-slope and prodelta in northeast Finnmark, North Norway – A possible ancient upper continental slope. *Jl. Sed. Petrol.* **52,** 171-186.

Pickering, K.T. 1984. Facies, facies association and sediment transport /deposition processes in a late Precambrian upper basin- slope/prodelta, Finnmark, N. Norway. *In:* **Stow, D.V.A. & Piper, D.J.W. (eds.).** *Fine grained sediments: Deep-Water processes and Facies.* Jl. geol. Soc. Lond. Spec. Publ. **15,** 343-362.

Plimmer, I.R. 1983. The association of tourmaline-bearing rocks with mineralisation at Broken Hill, NSW. *Aus. I. M. M. Conf. Broken Hill, NSW.* **July 1983.** 157 176.

Prior, D.B., Bornhold, B.D., Johns, M.W. 1984. Depositional chracteristics of a submarine debris flow. *Jl. Geol.* **92,** 707-727.

Pringle, J. 1911. Note on the 'Lower Tremadoc' rocks of St. Davids. *Geol. Mag.* **8,** 556-571.

Pringle, J. 1930. The geology of Ramsey Island (Pembrokeshire). *Proc. geol. Assoc. Lond.* **41,**1-24.

Uitterdijk-Appel, P.W. 1984. Tourmaline in the early archaean Isua Supracrystal belt. West Greenland. *Jl. Geol.* **92,** 599-605.

Raaf, J.F.M.DE, Boersma, J.R. & van Gelder, A. 1977. Wave-generated structures and sequences for a shallow marine succession, Lower Carboniferous, Co. Cork, Ireland. *Sedimentology.* **24,** 451-483.

Reed, F.R.C. 1895. The geology of the country around Fishguard, Pembrokeshire. *Q. Jl. geol. Soc. Lond.* **51,** 149-194.

Roach, R.A. 1969. The composite nature of the St. Davids Head and Carn Lliddi intrusions of North Pembrokeshire. *In:* **Wood A. (ed).** *The Precambrian and Lower Palaeozoic Rocks of Wales.* Univ. Wales. Press. Cardiff. 409-433.

Roberts, B. 1981. Low-grade and very low-grade metabasic Ordovician rocks of the Llyn and Snowdonia, Gwynedd, North Wales. *Geol. Mag.* **118,** 189-200

Roberts, B. & Merriman, R.J. 1985. The distiction between burial and regional metamorphism in metapelites from North Wales: an analysis of isocryst patterns. *Jl. geol. Soc. Lond.* **142,** 615-624.

Robinson, D. & Bevins, R.E. 1986. Incipient metamorphism in the Lower Palaeozoic marginal basin of Wales. *Jl. metamorphic. Geol.* **4,** 101-113.

Robinson, D., Nicholls, R.A. & Thomas, L.J. 1980. Clay mineral evidence for low-grade Caledonian and Variscan metamorphism in south-western Dyfed, South Wales. *Min. Mag. Lond.* **43,** 857-863.

Rock, N.M.S. & Leake, B.E. 1984. The international mineralogical association amphibole nomenclature scheme: computerisation and its consequences. *Min. Mag. Lond.* **48,** 211-227.

Rowbotham, G. & Bevins, R.E. 1978. Textures and compostion of pyroxenes in basic igneous rocks of Ordovician age from North Pembrokeshire. *Jl. geol. Soc. Lond.* **135,** 589-590.

Rushton, A.W.A. 1974. The Cambrian of Wales and England. *In:* **Holland, C.H. (ed.).** *The Cambrian of the British Isles, Norden and Spitsbergen (Lower Paleozoic rocks of the world, 2).* Wiley & Sons. Lond. 43-121.

Sato, K. 1975. Stilpnomelane from the Fujigatani skarn ore deposit, Japan, and its bearing on the problem of ferric and ferrous stilpnomelanes. *Neues. Jabrbuch. fur. Mineralogie. Montashefte.* **4,** 179-193.

Savage, N.M. & Basset, M.G. 1985. Caradoc-Ashgill conodont faunas from Wales and the Welsh Borderland. *Palaeoentology.* **28,** 679-713.

Schiffman, P. & Lofgren, G.E. 1982. Dynamic crystallization studies on the Grande Ronde Pillow Basalts, Central Washington. *J. Geol.* **90,** 49-78.

Schiffman, P., Bird, D.K. & Elders, W.A. 1985. Hydrothermal mineralogy of calcareous sandstones from the Colorado River delta in the Cerro Preito geothermal system, Baja California, Mexico. *Min. Mag. Lond.* **49,** 435-449.

Schminke, H-U. 1967. Fused tuffs and pépérites in South-Central Washington. *Geol. Soc. Am. Bull.* **78,** 319-330.

Schmincke, H-U, Rautesschlien, M., Robinson, R.T. & Mehegan, J.M. 1983. Troodos extrusive series of Cyprus: A comparison with oceanic crust. *Geology.* **11,** 405-409.

Schuiling, R.D. & Vink, B.W. 1967. Stability relations of some titanium minerals (sphene, peroviskite, rutile, anatase.). *Geochem. Cosmochim. Acta.* **31,** 2399-2411.

Scott, P.W. 1980. Zoned pyroxenes and amphiboles from camptonites near Gran, Oslo region, Norway. *Min. Mag. Lond.* **43,** 913-917.

Seilacher, A. 1967. Bathymetry of trace fossils. *Mar. Geol.* **5,** 413-428

Seilacher, A. 1978. Use of trace fossil assemblages for recognising depositional environments. *Soc. Econ. Palaeo. Mineral.Short Caorse notes.* 185-200.

Seyfried, W.E. Jr. & Bischoff, J.L. 1977. Hydrothermal transport of heavy metals by seawater. The role of seawater-basalt ratios. *Earth. Planet. Sci. Letters.* **41,** 265-276.

Sievell, W.J. & Waterhouse, J.B. 1984. Ocean ridge metamorphism of the Patuki Volcanics, D'Urville Island, New Zealand. *Lithos.* **17,** 19-36.

Smith, R.E. & Smith, S.E. 1976. Comments on the use of Ti, Zr, Y, Sr, P and Nb in classification of basaltic magmas. *Earth. Planet. Sci. Letts.* **32,** 114-120.

Smith, R.E., Perdeix. J.L. & Parks, T.C. 1982. Burial metamorphism in the Hamersly Basin, Western Australia. *J. Petrol.* **23,** 75-102.

Soegaard, K. & Eriksson, K.A. 1985. Evidence of tide, storm, and wave interaction on a Precambrian siliciclastic shelf: the 1,700 M.Y. Ortega Group, New Mexico. *Jl. Sed. Petrol.* **55,** 672-684.

Sparks, R.S.J. & Wilson, C.J.N. 1983. Flow-head deposits in ash turbidites. *Geology.* **11,** 348-351.

Spooner, E.T.C. & Fyfe, W.S. 1973. Sub-sea-floor metamorphism, heat and mass transfer. *Contrib. mineral. Petrol.* **42,** 287-304.

Stanley, D.G. 1981. Unifites: structureless muds of gravity flow origin in Mediterranean Basins. *Geo. Marine. Letters.* **1,** 77-83

Staudigel, H. & Schminche, H-U. 1984. The Pliocene seamount series of La Palma/Canary Islands.*J. Geophys. Res.* **89,** 195-215.

Stow, D.A.V. & Shanmugam, G. 1980. Sequence of structures in fine-grained turbidites: comparison of recent deep-sea and ancient flysch sediments. *Sedimentary Geol.* **25,** 23-42.

Stow, D.A.V., Alam, M. & Piper, D.J.W. 1984. Sedimentology of the Halifax Formation, Nova Scotia:Lower Palaeozoic fine-grained turbidites. *In:* **Stow, D.V.A. & Piper, D.J.W. (eds.).** *Fine grained sediments: Deep-Water processes and Facies.* Jl. geol. Soc. Lond. Spec. Publ. **15,** 127-144.

Stow, D.A.V. & Piper, D.J.W. 1984. Deep water fine grained sediments; facies models *In:* **Stow, D.V.A. & Piper, D.J.W. (eds.).** *Fine grained sediments: Deep-Water processes and Facies.* Jl. geol. Soc. Lond. Spec. Publ. **15,** 611-646.

Stubblefield, C.J. 1956. Cambrian Palaeography in Britain.*In:* **Rodgers, J. (ed.).** *El sistema Camrico su palaeogeografia y el problema de su base.* Int. geol. Congr. XX Sess. Mexico, 1956. 1-43.

Surdam, R.C. 1969. Electron micropprope study of prehnite and pumpellyite from the Karmusten Group, Vancouver Island, British Columbia. *Am. Mineral.* **54,** 256-266.

Surdam, R.C. & Sheppard, R.A. 1978. Zeolites in saline, alkaline lake deposits. *In:* **Sand, L.B. & Mumpton, F.A. (eds.).** *Natural Zeolites: ocurrance, properties, use.* Permagon Press. Oxford. 145-174.

Swift, D.J.P., Figueiredo, A.G., Freeland, G.L. & Oertel, G.F. 1983. Hummocky cross-stratification and megaripples: a geological double standard. *Jl. Sed. Petrol.* **53,** 1295-1317.

Swift, D.J.P., Hudelson, P.M., Brenner, R.L. & Thompson, P. 1987. Shelf construction in a foreland basin: strom beds, shelf sandbodies, and shelf-slope depostional sequences in the Upper Createaceous Mesaverde Group, Book Cliffs Utah. *Sedimentology.* **34,** 423-457.

Thomas, A.T., Owens, R.M. & Ruston, A.W.A. 1984. *Trilobites in British Stratigraphy.* Geol. Soc. Lond. Spec. Rep. **16,** 78pp.

Thomas, G.E. & Thomas, T.M. 1956. The volcanic rocks of the area between Fishguard and Strumble Head, Pembrokeshire. *Q. Jl. geol. Soc.* **112,** 291-314.

Thomas, H.H. & Jones, O.T. 1912. On the Precambrian and Cambrian rocks of Brawdy, Haycastle, and Brimaston. *Q. J. geol. Soc. Lond.* **68,** 374-386.

Thomas, H.H. & Cox, A.H. 1924. The volcanic series of Trefgarn, Roch and Ambleston (Pembrokeshire). *Q. Jl. geol. Soc. Lond.* **122,** 45-62.

Thompson, A.B. 1971. P_{CO2} in low grade metamorphism: zeolite, carbonate, clay mineral, prehnite relations in the system $CaO\text{-}Al_2O_3\text{-}SiO_2\text{-}CO_2\text{-}H_2O$. *Contrib. mineral. Petrol.* **33,** 145-161.

Thorpe, R.S. 1979. Late Precambrian igneous activity in Southern Britain. *In:* **Harris, A.L., Holland, C.H. & Leake, B.E. (eds.).** *The Calodonides of the British Isles reviewed.* Spec. Publ. geol. Soc. Lond. **8,** 579-584.

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 plate tectonic evolution of England and Wales. *J. geol. Soc. Lond.* **141,** 512-536.

Toghill, P. 1970. A fauna from the Hendre Shales (Llandeilo) of the Mydrim area, Carmathernshire. *Proc. geol. Assoc. Lond. geol. Soc.* **163,** 121-129.

Turner, F.J. 1981. *Metamorphic Petrology: Mineralogical, Field, and Tectonic aspects.* McGraw-Hill International Series in the Earth and Planetery Sciences. 2nd edition.

Turner, P. 1977. Notes on the depositonal environment of the *Lingula* Flags in Dyfed, South Wales. *Proc. Yorks. geol. Soc.* **41,** 199-202.

Turner, P. 1979. Diagenetic origin of Cambrian marine red beds: Caerfai Bay Shales, Dyfed, South Wales. *Sed. Geol.* **24,** 269-281.

Trzcienski, W.E. Jr. & Birkett, T.C. 1982. Compositional variation of pumpellyite along the western margin of the Quebec Appalachians.*Can. Mineral.* **20,** 203-209.

Uitterdijk Appel, P.W. 1984. Tourmaline in the early Archaean Isua Supra-crustal Belt, west Greenland. *Jl. Geol.* **92,** 599-605.

Vuagnat, M. 1975. Pillows lava flows: isolated sacks or connected tubes ?. *Bull. Volcanol.* **39,** 581-589.

Vernon, R.M. 1976. *Metamorphic processes: Reactions and microstructure development.* George, Allen & Unwin, 247pp.

Vierick, L.G., Griffin, B.J., Schminke, H-U. & Pritchard. 1982. Volcanoclastic rocks of the Reydarfjordur Drill Hole, eastern Iceland. 2. Alteration. *J. Geophys. Res.* **87,** 6459-6476.

Visser, J.N.J. 1983 Submarine debris flow deposits from the Upper Carboniferous Dwyka Tillite Formation in the Kalahari Basin, South Africa. *Sedimentology.* **30,** 511-523

Wells, G., Bryan, W.B. & Pearce, T.H. 1979. Comparative morphology of ancient and modern pillow lavas. *Jl. Geol.* **87,** 427-440.

Wetzel, A. 1984. Bioturbation in deep-sea fine-grained sediments: influence of sediment texture, turbidite frequency and rates of environmental change. *In:* Stow, D.V.A. & Piper, D.J.W. (eds.). *Fine grained sediments: Deep-Water processes and Facies.* Jl. geol. Soc. Lond. Spec. Publ. **15,** 595-610.

Whittington, H.B., Dean, W.T., Fortey, R.A., Rickards, R.B., Rushton, A.W.A. & Wright, A.D. 1984. Definition of the Tremadoc Series and Series of the Ordovician System in Britain. *Geol. Mag.* **12,** 17-33.

Williams A. *et al.* 1972. A correlation of Ordovician rocks in the British Isles. *Geol. Soc. Lond. Spec Rep.* **3,** 77pp

Williams, B.P.J. & Stead. J.T. 1982. The Cambrian rocks of the Newgale - St. Davids area. In: Basset, M.G. (ed). *Geological excursions in Dyfed, south-west Wales.* National Museum of Wales, Cardiff. pp. 27-49.

Wright, J.V. & Cas. R.A.F. 1986. Critical assessment of the concept of subaqueous pyroclastic flows and ignimbrites. *International Volcanological Congress 1986, New Zealand, Abs.Volume.* p.89.

Wybrecht, E., Duplay, J., Piqué, A. & Weber, K. 1985. Mineralogical and chemical evolution of White micas and chlorites, from diagenesis to low-grade metamorphism: data from various size fractions of greywackes (Middle Cambrian, Morocco). *Min. Mag. Lond.* **49,** 401-411.

Yamada, E. 1984. Subaqueous pyroclastic flows: their development and their deposits. *In:* **Kokelaar, B.P. & Howells, M.F. (eds.).** *Marginal Basin Geology; Volcanic and Associated Sedimentary and Tectonic Processes in Modern and Ancient Marginal Basins.* Spec. Publ. geol. Soc. Lond. **16,** 29-35.

Yamagishi, H. 1985. Growth of pillow lobes - evidence for pillow lavas of Hokkaido, Japan, and North Island, New Zealand. *Geology.* **13,** 499-502.

Zalasiewicz, J.A. 1984. A re-examination of the type Arenig Series. *Geol. Jl.* **19,** 105-124.

Zen, E-An. & Thompson, A. B. 1974. Low-grade regional metamorphism: mineral equilibrium relationships. *Ann. Rev. Earth. Planet. Sci.* **2,** 179-212.