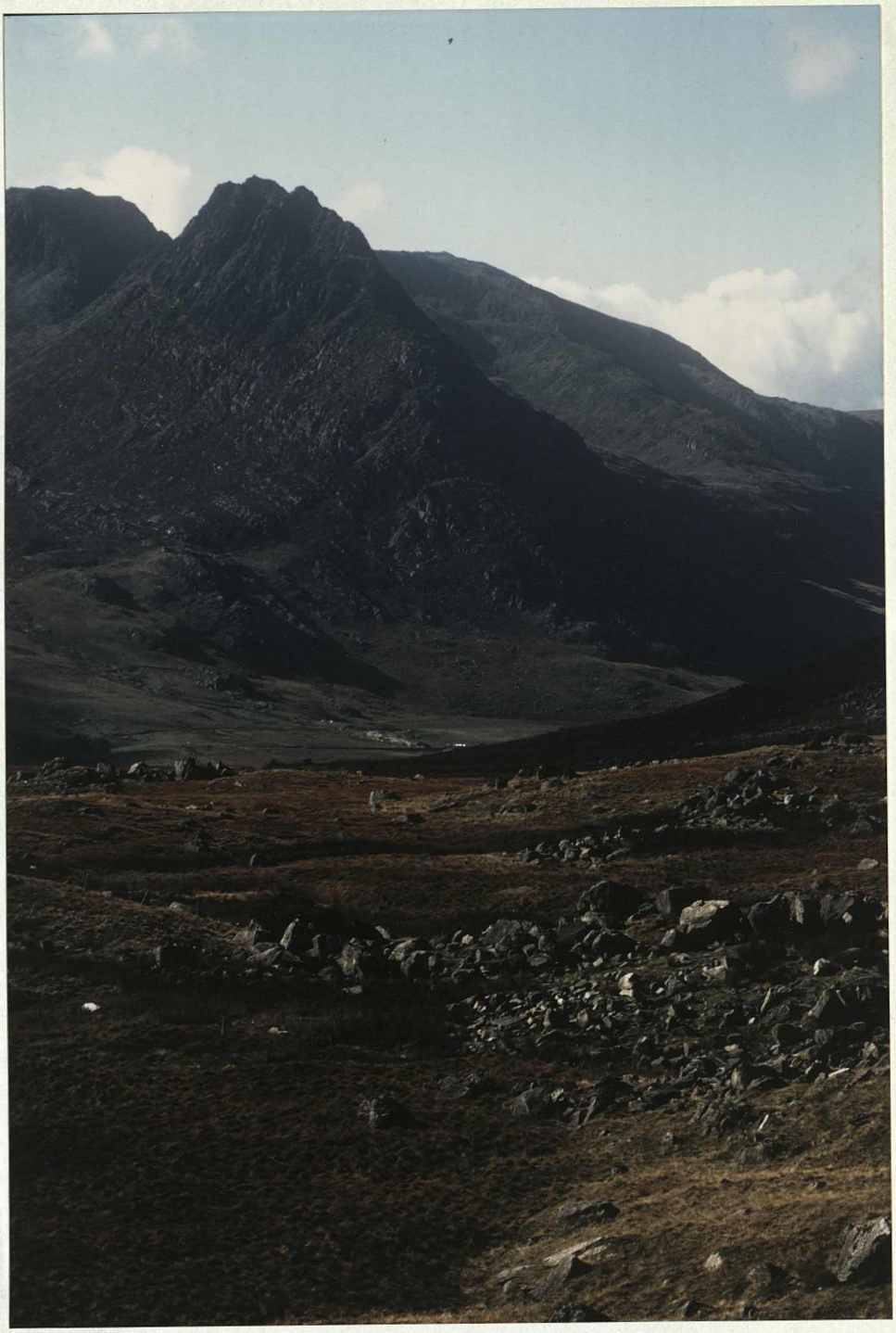

" Let us settle ourselves, and work and wedge our feet downwards through the mud and slush of opinion, and prejudice, and tradition, and delusion, and appearance, that alluvium which covers the globe, through Paris and London, through New York and Boston and Concord, through church and state, through poetry and philosophy and religion, till we come to a hard bottom and rocks in place, which we can call reality and say this is Truth."

Thoreau, 1854, Walden,

VOLCANICLASTIC SEDIMENTATION IN A CARADOCIAN MARGINAL BASIN, NORTH WALES

D. Phil. thesis
Department of Earth Sciences
University of Oxford

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VOLCANICLASTIC SEDIMENTATION IN A CARADOCIAN MARGINAL BASIN, NORTH WALES

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Wolfson College

Thesis submitted for the degree of Doctor of Philosophy
Trinity, 1988

ABSTRACT

A spectrum of volcaniclastic sequences have been recognised in the infill of a tectonically-active Caradocian ensialic marginal basin in North Wales. Volcanism within the basin was bimodal with extrusion of both rhyolite and basalt. Pyroclastic eruptions, usually rhyolitic, were extremely energetic with long recurrence intervals. Large volcanic events, in producing welded sheets of ash flow tuff obliterated pre-existing source terrains; any loose volcanic ash resulting from smaller pyroclastic eruptions was contained within localised volcanic cones and had little effect on the surrounding sedimentary budget. Depositional facies of both eruption styles suggest little intrabasinal volcanic relief. Correlation of the more laterally extensive units using geochemistry provides a useful stratigraphic framework for examination of the intercalated epiclastic successions.

Several hundred metres of such sediment were deposited in a wide variety of environments. Lines of basement weakness were repeatedly activated as faults, controlling both basin configuration and alluvial catchment areas. A variety of coarse-grained fan- and braidplain-delta types are distinguished, controlled by (1) the nature and geometry of fluvial depositional systems, and (2) the orientation of coastlines with respect to directional variation in the intensity of wave energy flux. Sandstone petrography indicates most delta systems derived their sediment load from erosion of uplifted granitic and sedimentary fault blocks with a subordinate contribution from coeval volcanism.

Integration of volcanological and sedimentary data shows that volcanotectonic uplift and subsidence was the dominant control on sedimentation, with the most significant changes in basin palaeogeogeography related to silicic eruptions. The magnitude and style of both volcanic events and associated tectonic changes is related to variation in crustal strength resulting from the emplacement, migration and vesiculation of magma bodies.

To my parents,

for their support and encouragement

over the years.

EXTENDED ABSTRACT

VOLCANICLASTIC SEDIMENTATION IN A CARADOCIAN MARGINAL BASIN, NORTH WALES

Geoff Orton
Wolfson College, Oxford

North Wales, in the mid-Ordovician or Caradoc, lay within an ensialic marginal basin. Volcanism within the basin was essentially bimodal, with voluminous eruptions of both rhyolites and tholeitic basalts. Basalts display ocean floor affinities with transitions to arc-like compositions; crustal extension is indicated by magma composition as well as the style of volcanicity and sedimentation. Rifting, however, was insufficient to allow production of oceanic crust and the basin developed on immature continental crust. Pronounced structural weaknesses within this basement were repeatedly reactivated as faults controlling both basin configuration and the resultant distribution of volcanic and sedimentary successions.

Research to date in North Wales has concentrated on mapping, lithostratigraphy, a genetic interpretation for the primary volcanic/volcaniclastic units and igneous geochemistry. The sedimentary component has been largely overlooked. The basin thus provides an ideal case study to examine the complex interplay between sedimentation, volcanicity and tectonics in an extensional basin. The objectives of research in North Wales and the principal aims of this thesis are as follows:

1). reconstruction of depositional environments of primary welded ash flow tuffs; to document the style of disruption of ash flow tuffs upon entry into the sea.

2). determination of the influence of volcanism on both sedimentation and tectonics within the basin and vice versa.

3). interpretation of the style of volcaniclastic sedimentation within an extensional ensialic basin.

Sequences have been variably deformed and metamorphosed by Caledonian movements (?early to mid Devonian; Shackleton, 1954; George, 1961) with metamorphism reaching lower greenschist facies in Snowdonia (Robinson & Bevins, 1985). Although the pervasive cleavage obliterates most sedimentary structures in lithologies finer than coarse siltstone (unless siliceous) and thus restricts facies analysis to sandstones or conglomerates, the folding and buckling has resulted in a 3-dimensional repetition of a 3km thick stratigraphic pile within a topographic relief

of less than 1000m. Measurement of numerous sedimentary sections within this framework forms the bulk of data gathered for the thesis and allowed an effective reconstruction of depositional systems and basin palaeogeographies with time. Facies analysis of depositional successions was supplemented by both petrography and geochemistry where necessary.

CHAPTER 2: STRATIGRAPHY: BEYOND THE LAYER CAKE

The extremely rapid changes in grain size and facies displayed by these sediments, the subtle interfingering of depositional systems from coeval volcanic and sedimentary sources, complicated by concurrent fault movements, has resulted in a complex assemblage of locally mappable lensoid units. The application of standard **biostratigraphic** or **lithostratigraphic** techniques, has generally met with failure. The endemism of shelly fauna and problems of their relationship with graptolite biozones at most levels, coupled with sparse biostratigraphic data from conodonts and other taxa, have hampered biostratigraphic correlations and erection of an acceptable chronostratigraphy (see Bassett, 1984). The application of lithostratigraphy to North Wales, although more successful than any biostratigraphic approach, has not been spectacular.

Tephrochronology was used to overcome the above problems. Over 100 samples were analysed for trace elements. Examination of the amount variation in whole-rock geochemistry within and between mapped ash flow tuffs showed that each tuff had a unique chemical signature. "Unidentified" volcanic units were classified into one of the known tuff groups; comparisons were verified statistically. The most important results obtained were:

1) correlation ash-flow tuffs such as the Pitts Head Tuff with their previously unrecognized distal equivalents and with co-ignmibrite air fall ashes.

2). recognition that the 2nd and 4th members of the Capel Curig Volcanic Formation were coeval in their eruption as suggested by Howells *et al*, 1980.

3). correlation of Glanrafon sandstones at Moel Hebog and Yr Arddu with Llewelyn Volcanic Group/ Cwm Eigiau Formation in NE Snowdonia.

CHAPTER 3: SANDSTONE PETROGRAPHY: THE VOLCANIC CONTRIBUTION

The volcanic contribution to epiclastic sedimentation in North Wales has long been recognized (*eg*. Ramsay,1866). In this chapter, the nature (*ie*. direct or indirect) and extent of the volcanic contribution to the

clasts of welded or even unwelded ash flow tuff among detritus was used to imply that major pyroclastic eruptions of rhyolitic composition were largely preserved in situ (*eg.* Garth tuff, Pitts Head Tuff) within lowland areas. High-level rhyolite intrusions, in contrast, appear to have cropped out in uplifted areas and provided the rhyolitic lithic fragments. The discovery of a hidden source of basaltic detritus within the basin has important implications for the structural evolution of the basin, importance of various fault trends in generating magma, and the tectonic stress regime in North Wales during the Caradoc.

The study area can be divided petrographically into two sub-basins about Llanberis pass, with input of basaltic and mudstone detritus into the SW portion, and a mixed source of tuffaceous sediment, rhyolitic volcanics and a granitic basement feeding the basin in the northeast (Fig. 3.128). For a large part of the evolution of the basin, these two depocentres were segregated by the central uplift of a small rhyolitic quartz porphyry horst block in the vicinity of Cwm Dyli. Besides influencing the petrography of local sediments, this topographic high guided the entry of the Pitts Head Tuff into the sea (Chapt. 7.3). In the NE sub-basin, there is a gradual reduction in the importance of uplifted granitic/sedimentary fault blocks as sources of sediment. The Upper Cwm Eigiau Fm. represents the culmination of this trend and records the most prolonged input of volcaniclastic material (both epiclastic and pyroclastic) into the basin. In spite of this, these sediments are among the most quartz-rich of all deposits examined. A review of literature shows that quartzose volcaniclastic sandstones are a common component of the infill of ensialic marginal basins or continental rifts. The implications of these findings on the validity of the recent plethora of ternary discriminant diagrams (*eg.* Dickinson and Suczek, 1979) purporting to provide information about the plate tectonic setting of depositional basins from sandstone petrography is discussed.

CHAPTER 4: FAN-DELTAS and BRAIDPLAIN DELTAS

Interpretation of coarse-grained fan- and braidplain-deltas, with associated coastline environments, has provided the most useful information regarding possible controls of volcanicity on basin tectonics and related sedimentation. In this chapter, 8 coarse-grained deltaic depositional systems are reconstructed. Emphasis is placed on description and analysis of distal alluvial plain and nearshore delta front deposits, where fluvial and marine interactions result in delicate

petrography of clastic sediments intercalated within known pyroclastic deposits is evaluated. Over 300 thin sections were examined petrographically, from both primary and secondary pyroclastic deposits, as well as volcaniclastic sandstones and siltstones, with 92 samples, representing all parts of the stratigraphy, selected for analysis by point-counting. Both standard optical and back-scatter SEM microscopy were used.

The information on the preservation of volcanic textures within reworked pyroclastics (Chapt. 3.2) allowed a genetic interpretation to be given to the white weathering "tuffites" which form an integral part of sequences within the Upper Cwm Eigiau Formation above the Pitts Head Tuff. The chemical and physical homogeneity of tuff beds, the thickness of the deposits, the complete lack of epiclastic material, and most importantly, the abundance of glass shards which do not survive marine reworking within secondary pyroclastics all point to a primary pyroclastic origin with NO storage of sediment before deposition. Each "tuffite" horizon is now interpreted to represent entry of vitric-rich pyroclastic flows into the sea. The difficulty in recognising volcanic material in thin section within either primary or secondary pyroclastics is discussed with reference to identification of chert in thin section.

The detrital composition of a sandstone can be related to several variables (1) provenance (*ie.* composition and relief of source area), (2) climate, (3) modification during transport and deposition and (4) alteration during diagenesis (Suttner, 1974). The influence of transport history and depositional environment was decreased through selection of samples from all portions of each depositional system with restriction to the alluvial component if possible. The role of grain size was reduced by including samples of available size fractions (from medium sand to granule conglomerate) in data sets for each depositional system. Climate was assumed to be an unchanging variable throughout the time span encountered. Sandstone petrography and depositional facies indicate that chemical weathering did not have time to significantly modify the original detrital mineralogy, and that the detrital signature was not modified beyond recognition by later diagenesis. In this light, the petrography of sandstones was used to monitor stratigraphic and regional changes in the nature of source rocks within the basin.

The influence of a granitic basement, uplifted fault blocks of tuffaceous sediment, and rhyolite volcanic/volcaniclastics was perhaps expected considering the known regional context. The paucity of obvious

and complex environmental differentiation. Fan deltas (4 examples) are distinguished from braidplain deltas (3 examples) and a 'common' delta based on the geometry and channel type of the contributing alluvial input. The latter two types of delta are basin axial systems fed laterally by alluvial fans. In the absence of any known variation in climate, the recognised diversity of alluvial systems (and resultant delta types) is attributed to the tectonic control on catchment areas.

Palaeogeographical configurations varied between deltaic systems, with major changes in sediment dispersal following volcanic eruptions and emplacement of voluminous ash-flow sheets (Orton, 1986; Chapt. 6). Since possible tidal deposits are rare within the studied successions, and where present suggest microtidal conditions (Chapt. 5.4.2), the spectrum of delta types primarily reflects a competition between fluvial and wave processes. Most deltas appear to have prograded into moderately deep basins, where wave energy was moderately high to extreme depending on coastline orientation. Largest wave ripples in all parts of the stratigraphy (Chapt. 5.5) indicate that maximum wind fetch was from the present south/southeast.

The inferred local variations in overall wave climate within the basin, however, appear rather insignificant when compared with the actual spectrum of delta types. The contrasting delta morphologies and associated patterns of sedimentation (summary in Fig. 4.82) are thus primarily a consequence of differences in fluvial processes. The critical parameters which controlled these deltas and determined their differing styles of development are thought to have been the geometry and slope of the alluvial "feeder" system, the rate of sediment supply (discharge magnitude and temporal variability), and the grain-size of sediment delivered to the delta front. These physical parameters varied among and between the alluvial fans and braidplains (see below), and dictated the varying responses of the associated deltas to basinal wave-climate.

To understand processes of nearshore sedimentation on a delta front, dissipative and reflective shoreline morphologies (Wright *et al*, 1979; Chapt. 5.3.5) were ascertained. Inferred dissipative coastlines had low overall gradients, a complex 3 dimensional topography with common longshore bars cut by rip currents. In contrast, inferred reflective beaches were be characterised by smooth 2-dimensional nearshore profiles with most wave energy reflected directly from the beach itself. Abundant fine sediment, a wide shallow inshore profile and high wave energy was found to produce dissipative conditions; a coarser grain size, reduced

sediment supply and lower wave power increased reflectiveness at the beach face. It is suggested the ternary classification diagram of common deltas should be extended in a fourth dimension to account for the grain size of sediment delivered to the delta front along the axis of the alluvio-deltaic prism. This makes it possible to compare deltas will similar gradients on the alluvial plain and delta front, similar responses to nearshore wave energy, and similar catchment areas and rates of clastic supply.**

CHAPTER 5: SHALLOW MARINE FACIES.

Shallow marine facies, with no obvious deltaic associations, form the bulk of Caradocian sedimentary successions in North Wales. Three shoreface successions reveal that fairweather and storm **wave** processes were the dominant factors responsible for redistribution of nearshore sediments. However, whereas fan- and braidplain-deltas commonly displayed evidence for **storm** processes (*eg.* wave ripples, HCS) within nearshore deposits, the non-deltaic shorelines, regardless of orientation, were dominated by fairweather wave and longshore processes. The contrasting facies associations are related to the varying preservation potential of storm events as determined by rates of sedimentation.

Shelf successions are examined within the Upper Cwm Eigiau Fm. comprising sediment between the Pitts Head Tuff and the Snowdon Volcanic Group. Profound changes in thickness are related to differing degrees of fault-controlled basin subsidence, with the Llanberis Pass, Idwal-Moel Hebog, and Nantmor-Llyn Cowlyd framework of fault systems being particularly important. In addition to allowing different rates of subsidence between adjacent fault blocks, the tectonic framework also controlled both depositional slope and the positioning of various depositional environments. Successions indicate a subtle interplay between sediment supply, rhyolite volcanicity, and basin tectonics. Variation on the theme within the basin is provided by different rates of tectonic subsidence, local topographic relief along related faults, and the composition of magmatic eruptions.

The shallow marine processes and inferred depositional environments were used to provide broad constraints on the bathymetric gradient. Calculation of wave parameters from available sand and gravel ripples using linear wave theory is also used to identify limiting conditions on both water depth and wave climate. This data confirms suggestions from depositional facies in deltaic successions (Chapt. 4.4.1) that wave power approaching from SSW to SSE was an order of magnitude larger than wave

energy approaching from any other compass direction. The ancient storm wave periods of 5-8 seconds implies an association with extratropical cyclones (gales) or winter storms rather than hurricanes; enhancement of SE trade winds appear to be the most viable explanation for their genesis.

CHAPTER 6: BASIN EVOLUTION: THE UPS and DOWNS

Changes in depositional enviroments, style of sedimentation and basin paleogeography are considered with respect to laterally extensive ash flow sheets or volcanic units. These are, with minimum depositional volumes, as follows:

Gwern Gof Tuff	2 km^3
Capel Curig Volcanic Fm: Garth tuff	28 km^3
Racks tuff	15 km^3
Pitts Head Tuff	70 km^3
Lower Rhyolitic Tuff	54 km^3

A strong control of volcanotectonic uplift and subsidence on basin evolution is illustrated. On a regional scale, before major eruptions, the emplacement and vesiculation of the rhyolitic magma bodies in mid-crustal levels resulted in the tilting of fault blocks, and as a result palaeoslope, away from the sites of magma introduction. More local changes in basin palaeogeography that always occurred at a stratigraphic distance of 20-30 metres beneath tuff horizons are thought to reflect the penultimate stages of magma emplacement into a fractured continental crust. There is a direct correlation between the amount of pre-eruption uplift and the size of individual volcanic events. The main pyroclastic eruptions and partial evacuation of magma chambers resulted in basin subsidence on both a local (*eg.* the Pitts Head tuff volcanotectonic trough, the Lower Rhyolitic Tuff caldera) and regional scale (above the Gwern Gof tuff, Garth tuff, the lower outflow sheet of the Pitts Head tuff). The amount and rate of post-eruption subsidence is again proportional to the magnitude of the preceeding volcanic eruption. It is suggested that the variance throughout the evolution of the Welsh basin in the sizes of ash flow tuffs and corresponding amounts of uplift can BOTH be related to the control of intrabasinal fracturing on continental strength and flexural rigidity.

CHAPTER 7: PRIMARY and SECONDARY PYROCLASTICS

As large ash flow sheets have been cited (Fisher & Schmincke, 1984, p. 293) as the best documented examples of subaqueous welded pyroclastic

flows, it is pertinent to reconsider this hypothesis in the context of the new interpretations of depositional environments and basin bathymetry (Chapt. 4-6), and identification of more distal correlatives of several pyroclastic units (Chapt 2.2.4.4). Ash flows in North Wales, which all must have had densities greater than 1 kg/m^3, all disintegrated upon entry in to the sea to either generate fine-grained phreatomagmatic ash, or a series of unwelded subaqueous debris flows. Sedimentary facies and brachiopod palaeocommunities indicate that the maximum depth of water in which welding occurred was 40 m. The contrasting patterns of flow disruption and style of mixing with seawater were related to inferred flow temperatures, flow velocities, depositional slopes, amounts of fluidization and related morphology of the flow front. Discussion focuses on the Garth and Pitts Head Tuffs.

The behaviour of the Pitts Head Tuff (PHT) upon entry into the sea varied with distance from its source. Differences are related to 1) inferred decrease in flow velocity further from source and related amount of fluidisation and 2) the depositional slope and tectonic activity accompanying flow emplacement. A pulsating flow front is used to explain a <u>series</u> of derivative sedimentary gravity flows resulting from flow transformations. Primary welding is invoked to account for textural features at the top of distal successions at Llyn Cwm-y-ffynnon.

The irregular base to the Garth tuff is related to transverse variation in shear stress along the front of a gravity current associated with lobe and cleft development. The spacing of the lobate protrusions are used to determine the amount of fluidisation of the flow during transport. Conflicting models for genesis of the Capel pod facies are unified by treating the Garth tuff as a fluidised, vigorously advancing, dynamic sediment gravity flow with a turbulent, polylobate flow front. Upon entry into the sea, the pyroclastic flow would start ingesting water into the base of the flow head, with preferential streaming of volatiles up through the entire flow along clefts. It is envisaged that this localised mixing pattern was maintained until the nose of the flow started to 'dive' under water. At this point, the additional incorporation of steam and water into the top and wake of the head through Kelvin-Helmoltz instabilities appears to have been 'too much for the flow to handle', and most of the Garth tuff was quickly transformed to a <u>dilute, water-supported</u>, sediment gravity flow. The Capel pods represent dense "hot spots" (perhaps plugs of primarily welded tuff) from the unmixed centre of the flow lobes.

Experimental evidence (Friedman *et al*, 1963; Shaw, 1963) on the viscosity of hydrated glass, and field evidence which suggests that the Welsh ash flows did incorporate large quantities of fluid, are used to suggest that "welding" textures cannot be related to plastic deformation of glass shards. Diagenetic pressure solution is proposed as a viable process to achieve the same end product in subaqueous settings. A preliminary review of literature suggests that development of welding is restricted to either extremely shallow water (less than flow thickness, *eg*. North Wales) from flows entering the sea or to deep water (greater than 500 m) where hydrostatic pressures confines submarine eruption columns and the increased solubility of water in glass decreases its viscosity (Sparks *et al*, 1980).

CHAPTER 8: SEDIMENTATION, TECTONICS and VOLCANICITY

A synthesis of the tectonic fabric which governed the formation, distribution, and preservation of the volcanic and sedimentary rocks, is used to illustrate that the axis of the basin in North Wales was actually aligned NW-SE (Chapt. 8.1). The dominant structural weaknesses which controlled basin evolution were the Conwy, Llanberis Pass and to a lesser extent Cwm Pennant fault systems. Preliminary results of an integration of available sedimentological, palaeontological, volcanic and tectonic data from all parts of the Avalonian microcontinent are presented to show that the Welsh marginal basin was probably related to SW subduction of the Tornquist under the Midland platform.

The different style of volcaniclastic sedimentation and basin fill within this extensional basin, when compared with convergent margin volcanic arcs, is related to variables such as magma chemistry, eruption size and frequency, grain size of pyroclasts produced and relative importance of intrabasinal volcanic and tectonic relief. Most volcaniclastic sedimentation in North Wales represents reworking of previously deposited pyroclastics by organised depositional systems during the long periods of time between volcanic events. Sandstone petrography indicates derivation of most detritus from uplifted fault blocks of volcaniclastic sediment, rhyolitic plutons, or granite rather than penecontemporaneous pyroclastics. The dominant control of volcanism on sedimentation was in causing numerous tectonic changes within the basin, with the most significant changes in basin palaeogeography related to silicic eruptions.

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The seed for this research project was originally planted by Malcolm Howells, who recognized that the complete absence of detailed sedimentology in a "classic" area of British Geology was detrimental to our understanding of not only the Welsh basin but volcaniclastic sedimentology. Thanks are due for his tempestuous collaboration, assistance with "finding" myself on the stratigraphy, cynical criticism of my sedimentology, and most importantly release of unpublished 1:10,000 base maps which proved invaluable. Other British Geological Survey colleagues, including Diarmid Campbell, Tony Reedman, and Martin Smith, freely offered time in the field when requested, periodically sent offprints before publication, and shared their personal insight into the complex interplay between tectonics, volcanicity and sedimentation.

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INTRODUCTION: OUT OF THE MISTS

> "I furnished the raw material, that material of which I had far too much, which I didn't know what to do with: existance, my existance"
>
> *Sartre, Nausea, p.143.*

1.1 VOLCANICLASTIC SEDIMENTATION

Volcaniclastic rocks, as broadly defined by Fisher(1961) are "the entire spectrum of fragmental volcanic rocks formed by any mechanism or origin, emplaced in any physiographic environment(on land, under water, or under ice) or mixed with any other clastic type". They have been genetically divided into four groups according to the method of fragmentation. These are pyroclastics (explosive disintegration of magma resulting from release of magmatic gases), autoclastic(fragmentation by mechanical friction during movement of lava or collapse of domes), hydroclastic(result of steam explosion at magma-water interfaces) and epiclastic(weathering and erosion of older volcanic rocks). In the last type, the dominant process of sedimentation is usually not a direct result of volcanicity, but reflects later reworking.

The abundance of data from the DSDP and recent investigations of island arcs (*eg.* Sigurdsson, Sparks et al, 1980) combined with earlier studies from ancient geosynclines (*eg.* Mitchell, 1970; Dickinson, 1974), have revealed the ubiquitous presence of all types of volcaniclastic rocks intermixed with non-volcanic debris within sedimentary basins through much of geological time. Garrels & MacKenzie(1971), based largely on the measured sequences of Ronov (1964), estimated that volcaniclastic rocks have formed 27% of post-Archean sedimentary rocks (Fig. 1.1). The contribution is probably even greater than shown, because of the amount of fine-grained tephra, altered to montmorillonite or illite, which remains unnoticed within shales (Fisher & Schmincke ,1984, p.8). Bjorlykke (1971,

1974) and Lindstrom (1974) have noted the volcanic contribution within Ordovician clays in particular. As Blatt (1985) notes that mudrocks comprise over 65% of the sedimentary stratigraphic column, an understanding of any potential volcanic contribution to a sedimentary basin is crucial in the interpretation of that basin.

Despite this mounting evidence for the importance of volcanic contribution, the study of volcaniclastics is still in its infancy. Largely stimulated by episodic volcanic eruptions such as that of Mt. St. Helens(1980), there has been some progress towards an understanding of primary volcaniclastics (see Fisher & Schmincke, 1984). There has been comparatively less advancement, however, in comprehending the common if not ubiquitous, epiclastic volcanic component of sedimentary sequences. Most volcanologists have focussed their attention on either igneous petrology, or pyroclastic processes; most sedimentologists have not even ventured near volcanically influenced regions, but concentrated on more conventional problems where facies models can be erected with greater ease.

Epiclastic volcaniclastics are notoriously difficult to interpret. The normal models of sedimentation used to interpret sediments in non-volcanic terrains no longer seem to apply (Reading, 1986; R. Suthren, 1988, pers. comm.). Some of these reasons are: (1) the diverse, compositionally immature and lithologically variable volcanic source rocks; (2) the wide range of hydrodynamic and aerodynamic properties of particles produced by this volcanicity; (3) the rapid volcano-tectonic movements that cause complex patterns of uplift and subsidence and rapidly changing water levels; (4) the episodic nature of sediment supply with repeated changes in depositional style related to extent and nature of volcanicity; and (5) the buildup of volcanic edifices, both subaerial and subaqueous, which can transform the topography and isolate depositional systems; (6) the local influence of volcanic eruptions on climate; (7) the destructive effects of

diagenesis. The above complexities have greatly hampered the study of volcaniclastic sedimentation. Most research has been from convergent margin volcanic arcs (*eg.* Guadeloupe, Mt. St. Helens), areas of high volcanic relief and andesitic volcanism (*eg.* Kuenzi, Horst *et al*, 1979; Mathisen & Vondra, 1983; Fritz & Harrison, 1985; Smith, 1986).

There have been even fewer detailed studies of volcaniclastic sedimentation from extensional basins. In the East African or Rio Grande rifts, for example, research has concentrated on detailed volcanic geochemistry, mapping, lithostratigraphy, and chemical sedimentation within lakes (*eg.* Frostick *et al*, 1986). Individual pyroclastic eruptions (often rhyolitic) in these continental rifts are more energetic with longer recurrence intervals and much larger volumes when compared with those from convergent margins. These silicic pyroclastic fields are also characterised by a gentle constructional topography and little volcanic relief (Lipman, 1984). Available evidence suggests a dramatically different style of volcaniclastic sedimentation (*eg.* Walton, 1986a, 1986b) in extensional basins when compared with convergent margin volcanic arcs.

North Wales is typical of extensional basins with a considerable volcanic component. Research has concentrated on mapping, lithostratigraphy and a genetic interpretation for the primary volcanic/volcaniclastic units. The sedimentary component has been largely overlooked. The basin thus provides an ideal case study to examine the complex interplay between volcaniclastic sedimentation, volcanicity and tectonics in an extensional basin. It has recently been noted (Barker, 1983, p.353) that such sedimentation is a more sensitive indicator of tectonic setting than the kinds and abundances of igneous rocks themselves.

1.2 HISTORY OF PREVIOUS RESEARCH

Wales has had a long and turbulent history (Fig. 1.2A). The astonishing thing about its recent history is that it can be written at all, for all history's cards seem to have been stacked against the survival of Wales as a separate entity. Its indestructible continuity was greatly aided by the existence of natural strongholds such as Snowdonia (frontispiece) into which a threatened people could periodically withdraw and safely rebuild their forces. These fortresses were determined by the geological structure of the country; this geology, in moulding the agriculture and industry, has continued to influence the course of Welsh history down to our day (Vaughan-Thomas, 1985).

If one plunges back another 450 Ma into Welsh history, to Caradoc times, one similarly finds a turbulent epoch with explosive pyroclastic events interrupting fragile periods of quiescence (Fig. 1.2b). This geological period reflects a complex interplay between volcanicity, sedimentation, and tectonics. The area of study (Fig.1.3, 1.4) extending from Conwy in the north, to Betws-y-Coed and Tremadoc in the south, covers an area of over 600 km^2, and encompasses the largest outpourings of Caradocian volcanics/volcaniclastics in Wales, from a number of sources. The intervening sediments provide a good record of the inherently complex tectonic evolution of the basin.

For over 150 years, North Wales has been an enigmatic attraction for geologists from many disciplines. A summary of research is best seen against a background of three time periods:

1. Early Mapping. 1831-1938.
2. The Welsh Geosyncline. 1938-1969.
3. An Ensialic Marginal Basin 1969-

More indepth reviews of developments during these periods, particularly in

other parts of the Welsh Basin, are given by Jones (1938), Bassett (1963), Bassett (1984) and Woodcock (1984).

1.2.1 EARLY MAPPING

From the earliest studies of the 19th century the cornerstone of geological investigations in Wales has been field mapping (Fig. 1.4) as a basis for a sound stratigraphy and accurate correlations (Bassett, 1984). The area formed the field ground for the long controversy between Murchison and Sedgwick over the stratigraphic subdivision of the Lower Palaeozoic (see Thackray, 1976).

Sedgwick, on his first visit to North Wales in 1831, determined the structure of the Snowdon Syncline, and the general succession of beds from the volcanic horizons of Snowdon itself to the rhyolitic rocks at Llyn Padarn. He noted (1847) that he knew "not a single instance (either in North or South Wales) of any unequivocal subaerial product. The series is essentially subaqueous".

In 1848, the systematic mapping of the area was commenced by the Geological Survey under the leadership of A.C. Ramsay; by 1852, both the Snowdon-Tremadoc and Bangor-Llanberis quarter sheets were published at a scale of 1 inch to 1 mile. In the accompanying North Wales memoir, Ramsay (1866) noted that a large amount of pyroclastic material occurred within the mass of the Snowdon 'felstones'. No attempt was made, however, to differentiate volcaniclastic units on geological maps as "sometimes they are pure and sometimes intermixed with every possible percentage of other sediments" (*op. cit.*, p.142). Ramsay also recognized that much of the "ashy material was recomposed, having just fallen on land around one or more of the volcanoes, from whence, mixed with ordinary sediments, it was spread abroad in the manner of other strata by the effects of marine denudation" (*op. cit.*, p.162). A 3-fold division of the volcanic rocks was established (Fig. 2.4) with a lower group of 'slaggy and brecciated

feldspathic porphyries' succeeded by 'sandy feldspathic and calcareous ashes' and this in turn by an upper columnar feldspathic trap; Ramsay stressed that this "Snowdon Volcanic Group" was not stratigraphically equivalent to the older volcanics of the Arenig, Aran and Moelwyn mountains. In referring to the Upper Ordovician Denbighshire grits, he later (1881, p.285-286) suggested that "from the quantity of feldspathic grains that enter into their composition...[one can]..infer that part of the mountain region of North Wales between Cader Idris and Conwy then formed land, which in some degree contributed locally to make these strata by the waste of feldspathic lava and ashes".

The petrography of the volcanic rocks was first described by Ward (1875), who noted that "finely bedded ash when highly altered is indistinguishable in microscopic texture from undoubted felstone lava". Later workers (Rutley, 1881; Mugge, 1893) noted that the cleaved lavas of Llanberis Pass resemble devitrified obsidian and that the Pitts Head "rhyolite" at Rhyd Ddu had the character of fine rhyolitic tuffs.

Harker (1889) made the first attempt to distinguish between the characters of 'lavas' of different horizons in Caernarvonshire, suggested lateral limits for successive volcanic suites and even identified possible volcanic centres. He felt that "fragmental and quasifragmental rocks of the acid volcanic series were much less extensive" than supposed by Ramsay. Lavas were presumed to have "solidified under the pressure of some depth of water". The alternative, which he dismissed would require " a long succession of upheavals and depressions to account for interbedded sedimentary strata".

In 1900, Dakyns noted that in many places the lower Snowdon felstones, described as lavas by Ramsay, were probably felsitic tuff. In a later, much overlooked, examination (Dakyns & Greenly, 1905), the Lower Rhyolitic tuff was interpreted as masses of fine volcanic dust rapidly deposited in a

manner similar to that of the ash from the dust cloud eruption of Mont Pelée in 1902. They also noted that although "it is true that the phenomena which have given some a dramatic interest in these eruptions.. ..are essentially subaerial, and the volcanic series of Snowdon is generally regarded as submarine", the conditions for Pelean eruptions "are determined in the throat of the volcano before emergence, there is nothing to prevent some eruptions taking place from submarine vents" (p. 548).

The most detailed mapping in North Wales during this early period was accomplished by the Williams brothers, Howel and David, who placed special emphasis on interpretation of the igneous activity. In 1922, Howel Williams, in mapping the volcanic units near Capel Curig, discriminated between vitric, crystal and lithic rich tuffs, noting that all types pass into one another gradually. He also noted the importance of the amount of water mixed with magma in controlling eruption style and resultant grain size associated with eruptions of basaltic tuffs at Capel Hill (SH723580). In his classic paper on the geology of Snowdon, Williams (1927) provided the first detailed description of the age and succession of both sedimentary and volcanic rocks of Snowdon. He later extended his mapping to the Dolwyddelan syncline (Williams & Bulman, 1931). A more complete stratigraphic section allowed correlation of black slates with similar deposits (Upper Cadnant black shales, Elles, 1909, p. 172) at Conwy mountain and constraint of the age of the Snowdon Volcanic Group and associated sediments to the Middle Caradoc. As these black slates immediately succeed a series of shallow water pyroclastic rocks, Williams suggested that it is unlikely that the slates were laid down at any great depth. He suggested deposition in basins where submarine barriers created condition of severe stagnation, a viewpoint later supported by Cave(1965) in his analysis of the Nod Glas shales. David Williams (1930) extended the mapping of his brother further to the northeast to Ogwen valley;

Davies(1936) provided the connecting link between Ordovician successions of Snowdonia and Conwy, and outlined the stratigraphy of the Crafnant Volcanic succession.

1.2.2. THE WELSH GEOSYNCLINE

The application by O.T. Jones (1938, 1956) of the geosynclinal concept to the Lower Palaeozoic rocks of Wales (Fig. 1.5), caused a revolution in thinking towards the Welsh Basin. The concept was quickly seized upon by a burgeoning band of sedimentologists. This new phase of vigorous research that resulted placed appreciably more emphasis on both the interpretation of depositional environments and the relationship between tectonics and sedimentation.

Jones originally separated lithofacies into two major categories (shelf and basinal) and outlined the evolution of the Welsh geosyncline based on the distribution of these two facies. He later(Jones and Pugh, 1949) described an extremely complicated pattern of volcanic and sedimentary facies interpreted as deltaic, estuarine and shallow water sediments, associated with a Llandeilo (Ordovician) volcanic complex at Builth Wells.

Most sedimentary research during this time, however, revolved around application of the turbidity current hypothesis to Cambrian or Silurian strata. Starting with the seminal work of Kuenen (*eg.* 1953), the recognition that some Lower Palaeozoic basinal grits and greywackes could have been deposited in deep water as turbidites, rapidly altered existing bathymetric models. Transport vectors from sole marks (*eg.* Wood and Smith, 1959; Cummins, 1957, 1959; Holland, 1959) and a plethora of data from slumped beds (*eg.* Jones, 1940; Boswell, 1940) allowed significant modification of ideas on basin palaeogeography.

Many workers, however, assumed all greywackes were transported by turbidity currents (*eg.* Cummins, 1957; Kuenen & Migliorini, 1950) and

Pettijohn (1950) stated "greywackes are marked by a primary mud matrix and are graded by reason of their deposition from turbidity currents". This linkage of the term greywacke with turbidity currents, rather than with sedimentary petrography as originally defined (Pettijohn *et al*, 1973, p. 159) and the related association of greywackes with geosynclines, created much confusion in the interpretation of the Welsh basin. In the absence of any sedimentary information from Ordovician sequences in North Wales, the view that nearly all of the Lower Palaeozoic rocks of Wales must be submarine because they were deposited in a geosyncline, persisted until the late 1950's. Mitchell (1957), for example, concluded that the majority of volcanic rocks were deposited under water, and that the tuffs and lavas of Rhobell Fawr were the only appreciable subaerial volcano in Wales.

A frenzied outburst of petrographic examinations, however, was soon to question the validity of this simplistic perception. Oliver(1954) had already demonstrated the unquestionable occurrence of welded texture within the Ordovician volcanics of Snowdonia. Subsequently Rast, Beavon & Fitch, in a series of papers(Beavon *et al*, 1961; Rast, 1960, 1961; Rast *et al*, 1958,1959) summarised by Fitch (1967) recognised that many units previously interpreted as rhyolite lavas, such as the Pitts Head Tuff and basal breccia of the Lower Rhyolitic Tuff were in fact welded ash flow tuffs. Brenchley (1964) similarly recognised the occurrence of welded tuffs within Caradocian strata of the Berwyn Hills and provided detailed descriptions of their petrography.

With the belief that welded tuffs could not form subaqueously emplacement, these discoveries led to substantial revision of palaeogeographical models for the Welsh Ordovician. The recognition of probable subaerial landmasses within the Welsh basin also revitalized queries regarding conditions of deposition of the black graptolitic shales, previously mentioned by Williams & Bulman, (1931). Authors usually invoked

the association of tectonism with volcanicity to account for the interbedded fossiliferous sediments. Rast (1961, p.646), for example, suggested that volcanotectonic uplift from viscous rhyolitic magma 'blisters' repeatedly created widespread areas of dry land before major ignimbrite eruptions. Volcanotectonic collapse was later appealed to by several authors. Beavon(1963), used it to explain the origin of the Llyn Dinas Breccia at the base of the Lower Rhyolitic Tuff east of the Glaslyn river; Roberts (1967, 1969) recognized that the Llwyd Mawr ignimbrite occupied a volcanotectonic collapse structure (p.353), which was genetically connected with the eruption and emplacement of the correlative Pitts Head Tuff flows.

Evidence either for or against these tectonic changes was limited. The petrography and sedimentary structures of intervening sandstones had received much less attention than those of the Cambrian or Silurian. Brenchley (1964) had made one of the earliest attempts to describe the sedimentology of Ordovician volcaniclastic rocks, with detailed analysis of sequences from the Berwyn Hills. In Snowdonia, however, the only hint of sedimentology by this time consisted of scattered observations by Shackleton (1959) and Beavon (1963) within more general papers. These authors noted that the coarse grain size, large wave ripples and common cross-bedding characteristic of Caradocian sediments (Gorllwyn grits) near Moel Hebog and Beddgelert, were indicative of a shallow marine environment.

1.2.3. THE ENSIALIC MARGINAL BASIN

Boswell (1961) argued cogently against use of the term geosyncline for the Welsh basin; he instead viewed the Lower Palaeozoic Welsh region as a series of small localised basins which display considerable variation in structure, lithofacies, and thickness and were modified at several stages by 'platform' tectonics. It was not, however, until plate tectonic theory

was applied to the development of geosynclines almost 10 years later that the full importance of Boswell's observations could be realised.

Following the original suggestion by Wilson (1966) for a proto-Atlantic ocean, later termed the 'Iapetus' by Harland and Gayer (1972), Dewey (1969) proposed the first model involving sea floor spreading and subduction zones to explain many features of the late Precambrian through early Palaeozoic geology of the Caledonides. The Welsh basin was depicted by Dewey (1969, his Fig.1, 2) as an intracratonic basin related to subduction of the oceanic lithosphere of the Iapetus. Fitton and Hughes (1970) subsequently used the geochemistry of volcanic rocks in the Lake District and Wales to support this hypothesis. Like Dewey, they invoked a second, more southerly Benioff zone to explain the earlier volcanic activity of Tremadoc-Arenig age at for example Rhobell Fawr. It was later suggested (Mitchell & Reading, 1971; Phillips *et al*, 1976) that Wales in the Ordovician developed as an ensialic basin formed by small scale back-arc spreading. Oblique subduction, and diachronous collision of converging continental margins were invoked (Phillips *et al*, 1976) to explain the ending of volcanism at progressively later dates toward the southwest. Later refinements to these models (Kokelaar, 1979; Dunkley, 1979; Kokelaar *et al*, 1984) recognised contrasting magma chemistries between the Tremadoc Rhobell Fawr volcanic complex and later Ordovician volcanics. The former display characteristics typical of island arc-destructive plate margin volcanism. The widespread Arenig transgression was interpreted to reflect a distinct change in tectonic conditions within the basin, heralding the development of the extensional Ordovician marginal basin. Woodcock (1984) has recently proposed that strike-slip movements were important in the development and evolution of the Welsh basin, particularly from Ashgill-times onward.

In a similar way to the earlier geosynclinal concept, the introduction of plate tectonic theory into the Welsh basin, has acted as a catalyst for new dynamic research in a wide variety of disciplines. In the last 10 years in particular, there has been an upsurge in the use of palaeontological and stratigraphic, structural, and sedimentological techniques.

Developments in Ordovician palaeontology/palaeoecology have not kept pace with those in other periods of the Lower Palaeozoic. Following Ziegler's(1965) recognition of regional patterns within lower Silurian marine paleocommunities, numerous authors have rigorously assessed relationships between sedimentary facies and brachiopod fauna, particularly from the type Ordovician (*eg.* Pickerill & Brenchley, 1979; Hurst, 1979b; Brenchley & Newall, 1980). Because the speciation of the dalmanellids and sowerbyellids is largely based on statistical methods (Williams, 1963), it is difficult to obtain satisfactory comparisons with type specimens, because of the deformed nature of the specimens and the common imperfect preservation (Diggens & Romano, 1968, p.603; Wright, 1979; Chapt.5). The only comprehensive examination of brachiopod fauna from the field area was by Wright (1979). He concluded that complex intergradation between shallow shelf communities, with community composition largely reflecting environmental stability and nature of the substrate, prevented recognition of any distinctive onshore-offshore patterns. The ground to the southeast of the field area between Betws-y-coed and Bala was mapped by Campbell (1983) who used a combination of volcanic and biostratigraphic horizons to subdivide the Caradoc stratigraphy, allowing broad correlation of some sequences with both the Berwyn Hills and Snowdonia.

The advancement in our understanding of the structural evolution of North Wales has similarly lagged behind that of other regions of the Welsh Basin. Shackleton (1954) broadly defined the morphology, orientation and

distribution of folds and other structural elements within North Wales; a simplified model for the tectonic evolution of the Welsh Caledonides was later proposed by Coward and Siddans(1979), based on little additional data. Only recently, however, have detailed structural studies commenced. Smith (1987, 1988), for example, has reinterpreted structural features of the basal Ordovician (previously described as the Tremadoc thrust zone) in terms of intrastratal gravity induced slumping; Wilkinson (1987,1988) has furnished a regional strain analysis of North Wales which suggests that structural deformation within fault blocks could be explained by a simple shear mechanism.

The application of sedimentology to lower Palaeozoic successions has continued to focus on the interpretation of the more "basinal" strata, with particular emphasis on the tectonic control of turbidite sedimentation within the Silurian of mid-Wales (Mackie & Smallwood, 1987; Dimberline & Woodcock, 1987; Tyler & Woodcock, 1987; Smith, 1987). James (1973, 1975, 1983) supplied detailed observations on Caradocian/Ashgill submarine fans from mid and south Wales. The recognition that storm processes can explain many facies associations has recently "shallowed" the Welsh basin. The common presence of hummocky cross-stratification, for example, has been noted from Caradoc age deposits in Shropshire and the Berwyn Hills of North Wales (Brenchley & Pickerill, 1980; Brenchley & Newall, 1982). A storm generated origin has also been invoked for "classical" Silurian "turbidites" of the Welsh Borders (Benton & Gray, 1981; Tyler & Woodcock, 1987) with evidence based on lateral heterogeneity within sequences, shelly bioclasts, and palaeocurrent trends. Woodcock (1984) concluded that many Welsh "turbidites" were deposited on an only modestly overdeepened continental shelf.

Recent sedimentological developments from Ordovician sequences in Snowdonia have been sparse, to say the least. The age and form of Lower to

Middle Llandeilo ironstones at Betws Garmon and ferruginous debris flows in Llanberis valley, have recently (Trythall *et al*, 1987) been described . Diggens and Romano (1968) outlined the stratigraphy and petrography of Caradocian sequences at Llyn Cowlyd (SH715615) and made crude environmental interpretations. Broad environmental descriptions were included within the papers by Howells and colleagues suggesting subaqueous welding of the ash flow tuffs (Francis & Howells, 1973; Howells & Leveridge, 1980; Howells *et al*, 1985). Although exposure precluded measurement of detailed sedimentary sections, Campbell's (1983) description of available sedimentary structures, combined with biostratigraphic and lithostratigraphic correlations, allowed the depositional history of Caradocian sequences to the southeast to be outlined. Here, most sandstones were deposited in shallow marine environments, often above storm wave base.

The emphasis of research in North Wales during the last quarter century, has continued to be placed on igneous and volcanic geology. As part of a substantial mapping project reexamining classical sequences in North Wales officers of the British Geological Survey (Francis & Howells, 1973; Howells et al, 1979) suggested that welding in ash flow tuffs can take place subaqueously as well as subaerially, so that there is no need to invoke repeated periods of emergence associated with their emplacement. Detailed mapping has allowed recognition of numerous volcanic centres, and variations within the composition, texture, and depositional character of ash flows were related to changing depositional environments (Howells et al, 1973, 1978, 1980, 1981a, 1981b, 1983; Howells & Leveridge, 1980). Campbell (1983) later mapped the distal equivalent of one of these supposed subaqueously welded units, the 1st member of the Capel Curig Volcanic Formation. The concept of subaqueous welding was poorly accepted at the time (see discussions in Francis & Howells, 1973) and most volcanologists

(e.g. Fisher & Schmincke, 1984, p.293-296; Cas & Wright, 1987, p. 276-280; S. Self, pers.comm., 1987) are today still sceptical.

Just as Kelling (1964) had earlier complained that the "dearth of detailed sedimentology from Ordovician rocks was an important contributory factor" in accounting for the apparent lack of turbidites within Ordovician successions, the lack of sedimentology has not allowed one to assess the validity of new volcanological models. Woodcock (1984) recognised that one of the outstanding sedimentological problems within the Welsh basin, was the interaction between volcanic and sedimentary processes.

1.3. SCOPE OF PRESENT STUDY

It was against this background, that sediments of the "highest, wildest, and grandest part of North Wales" (Ramsay, 1881), previously described as "very unpromising looking rocks"(Greenley, 1905, p.543) and thought of as 'inferior' to the Old Red sandstone (Sedgwick, 1838) were approached with caution. Sequences have been variably deformed and metamorphosed by Caledonian movements (?early to mid Devonian; Shackleton, 1954; George, 1961) with metamorphism reaching lower greenschist facies in Snowdonia (Robinson & Bevins, 1985). Resultant folds in Snowdonia (Fig. 1.6, 1.7a, b) define a broad arcuate trend; open folds with upright to SE dipping axial planes predominate but tight to isoclinal, locally overturned SE verging folds are also present. Structures often change from open to tight within 10 kilometres along individual fold hinges (Wilkinson, 1987). Thrusting does not seem to have been significant in North Wales, with any unconformities being syn-sedimentary in origin (Smith, 1987; this study). Bulk extension on cleavage associated with folding varies from 110% in rhyolitic tuffs, 120-140% within coarse sandstones and 170 to over 200% vertical extension (50% shortening) within mudstones (Coward & Siddans, 1979; Wilkinson, 1987). The intensity of cleavage is also closely related to lithology with slaty cleavage forming in mudstones (Fig. 1.8a) and

intergranular spaced fractures developing in coarser sediments (Fig. 1.8b). Although the pervasive cleavage obliterates most sedimentary structures in lithologies finer than coarse siltstone (unless siliceous) and thus restricts facies analysis to sandstones or conglomerates, the folding and buckling has resulted in a 3-dimensional repetition of a 3 km thick stratigraphic pile within a topographic relief of less than 1000 m. Measurement of numerous sedimentary sections within this framework allows an effective reconstruction of depositional systems and basin palaeogeographies with time.

The project originally proposed involved sedimentological examination of only the Cwm Eigiau Formation (Fig. 1.9) subjacent to the Snowdon and Crafnant Volcanic Groups in eastern and northeastern Snowdonia. After an initial reconnaissance, it quickly became clear that: (1) for full understanding of the interplay between volcanism and sedimentation the project needed to be stratigraphically widened to include sandstones/conglomerates subjacent to the Capel Curig Volcanic Formation within the Llewellyn Volcanic Group and (2) cleaved slates and mustones offered little useful sedimentary information.

During the first field season, numerous sections were measured in the Capel Curig and Tryfan anticlines, and Idwal syncline (Fig. 1.9). Preliminary palaeogeographic interpretations revealed frequent changes in the direction of basin fill throughout its history. As a result, the field area was extended laterally to Conwy in the NE and Tremadoc in the SW to examine more "proximal" sediments intercalated with the volcanic units (Fig. 1.9). Emphasis has been placed throughout research on measurement of detailed stratigraphic sections in the field, regardless of accessability or altitude. During the three field seasons over 8000 metres of stratigraphic section were measured at scales varying from 1:10 to 1:100 according to the amount of detail visible. Where possible, depositional

units were traced laterally between sections, or sections were measured at close (50m) lateral spacing, to enable the best reconstruction of depositional environments. The field area was divided for convenience into several regions of study (Fig. 1.9, 1.10); both vertical sections and petrographic samples are prefixed according to the area within which they fall. Logs from most vertical sections measured are included within Appendix 5 at a condensed scale.

Facies analysis of depositional successions was supplemented by both petrography and geochemistry. Major and trace element geochemistry from over 120 samples of rhyolitic pyroclastics was used to confirm lithostratigraphic correlations between ash flow tuffs and distal equivalents that were suggested by facies analysis of intercalated depositional systems. Petrographic examination of over 300 thin sections provided valuable information regarding 1) textures within primary pyroclastic units and 2) discrimination between episodes of deposition reflecting contemporaneous volcanicity and those resulting from uplift and erosion of previously deposited volcaniclastics, and 3) variation in sandstone source with time.

The objectives of research in North Wales, that have been slowly defined with time, and form the principal aims of this thesis are as follows:

1). reconstruction of depositional environments of primary welded ash flow tuffs; to document the mode of entry of ash flow tuffs into the sea (or not!!).

2). examination of the influence of volcanism on both sedimentation and tectonics within the basin and vice versa.

3). interpretation of the style of volcaniclastic sedimentation within an extensional ensialic basin.

1.4 HIERARCHY OF THESIS

In Chapter 2, the general tectonic setting and existing stratigraphy of North Wales is first outlined before tephrochronology is used to refine lithostratigraphic correlations. Successive chapters move progressively from smaller to larger scales of observation. At the lowest level, in Chapter 3, petrographic variation within sandstones is set against the refined stratigraphy with emphasis on determination of concurrent volcanicity and implications for source characteristics. Lateral and vertical variation within depositional sequences (Chapter 4, 5), in tandem with consideration of corresponding changes in biota, allows recognition of alluvial fan, deltaic, nearshore and shallow marine environments. At a larger scale, the sequential and geometric arrangement of vertical successions allows basinwide correlation of depositional systems and gives insight into tectonic changes associated within volcanicity (Chapt.6). The primary volcanic units are first discussed in Chapter 7; the previous reconstructions of depositional environments are used to comprehend the varying ability of ash-flow tuffs to enter the sea. Chapter 8 provides an integration of tectonic, sedimentary and volcanic information to comment on the style of volcaniclastic sedimentation in general. A brief guide to the pronunciation of Welsh is provided in Figure 1.11 to facilitate vocalization of place names for foreigners.

Chapter 2 STRATIGRAPHY

> "When I behold a rich landscape, it is less my purpose to recite correctly the order and superposition of the strata, than to know why all thought of multitude is lost in a tranquil sense of unity"
>
> *Emerson, 1836, Nature.*

2.1. INTRODUCTION

2.1.1 GENERAL SETTING

The Ordovician Period was characterized by extremes. It was a time of extensive volcanicity and tectonic unrest throughout the entire world (Fortey, 1984; Stillman, 1984; Williams, 1976), with most of the preserved volcanics erupted in island arcs, marginal basins, or active continental margins. In the British Isles, it also appears to have been a time of complex plate tectonic rotations (Piper, 1978) and/or an anomalous, fluctuating, geomagnetic field (Thomas & Briden, 1976). Epicontinental seas had a greater extent than in almost any other period. Extreme biogeographic differentiation produced by pronounced climatic zonation in the Southern Hemisphere(related to glaciation) was coupled with restricted communication between the various epicontinental seas; this is in direct contrast to the cosmopolitan faunal distribution of the succeeding Silurian epoch. As a result, stratigraphic classification of the Ordovician is perhaps more confused than any other Phanerozoic period (Janusson, 1984).

North Wales, in the mid-Ordovician or Caradoc, lay within an ensialic marginal basin (Kokelaar *et al*, 1984a). Although volcanism within the basin was essentially bimodal, with voluminous eruptions of both rhyolites and tholeitic basalts, minor calc-alkaline trachyandesites do occur. Basalts display ocean floor affinities with transitions to arc-like compositions: such characteristics are typical of marginal basin settings (Saunders & Tarney, 1984). Crustal extension is indicated by magma composition as well as the style of volcanicity and sedimentation.

Rifting, however, was insufficient to allow production of oceanic crust and the basin developed on immature continental crust (Watson & Dunning, 1979). Most basement appears younger than 900 Ma and probably developed by accretion of earlier volcanic arcs (Thorpe, 1974; Thorpe et al, 1984). Pronounced structural weaknesses within this basement were repeatedly reactivated as faults controlling both basin configuration and the resultant distribution of volcanic and sedimentary successions, as well as the style of later tectonic deformation (I. Wilkinson, pers. comm., 1987).

In North Wales, Ordovician volcanicity was confined to the Caradoc epoch, the rest of the Period being dominated by the deposition of mud (Fig.2.1). Arenig to early Caradoc strata, for example, comprise a monotonous sequence of marine mudstones, siltstones and impersistent sandstones (Howells *et al*, 1983); widespread black pelagic mudstones and muddy turbidites typify the Ashgill. The thick Caradocian succession (over 3km) has been divided into two major volcanic groups (Fig. 2.1): a lower Llewelyn Volcanic Group and an upper Snowdon Volcanic Group (Howells *et al*, 1978; Howells *et al*, 1983).

The exact age of these Caradocian volcanics is uncertain. The presence of *D. multidens* zone graptolites in slates underlying the Llewelyn Volcanic Group at Nant Ffrancon (SH 6463), and a Soudleyan shelly fauna in overlying sediments establish an early Caradoc age (Costonian to Soudleyan) for this lower volcanic succession (Howells *et al*, 1983). Bancroft (in Lamont, 1946) assigned a Costonian to Lower Soudleyan age to strata in the Tryfan anticline; a Soudleyan fauna has been obtained by several authors (Bancroft, *op. cit.*; Diggens & Romano, 1968) from the Capel Curig Volcanic Formation. The later Snowdon Volcanic Group seems to have been deposited entirely within the Longvillian. The base of the Lower Rhyolitic Tuff, where conformable, lies above the Soudleyan-Longvillian boundary and contains Lower Longvillian fauna (Harper, in Shackleton, 1959, p.249); the

Bedded Pyroclastics commonly contain Upper Longvillian fauna (Harper in Shackleton, 1959, p.250; Dean, 1956). Little sedimentation appears to have occurred in North Wales in the latest three stages of the Caradoc. The Soudleyan-Longvillian boundary has been presumed to coincide with the voluminous eruption of the Pitts Head Tuff (Howells *et al* , 1978; Campbell, 1983; Howells *et al* , 1986, p. 411) with Wright(1979, p.286) assigning the Pitts Head Tuff to the base of the Lower Longvillian. The validity of this assumption, however, is questioned later (Chapt. 2.2.2.4).

Primary rhyolitic ash flow tuffs, airfall tuffs, and pyroclastic breccias, as well as minor basalt lava flows and hyaloclastites, are intercalated with complex sequences of derived volcaniclastic sediment of comparable thickness. The epiclastic material varies in grain size from boulder conglomerate to mudstone. The extremely rapid changes in grain size and facies displayed by these sediments, the subtle interfingering of depositional systems from coeval volcanic and sedimentary sources, complicated by concurrent fault movements, has resulted in a complex assemblage of locally mappable lensoid units. The application of standard biostratigraphic or lithostratigraphic techniques, has generally met with failure, particularly within the epiclastic component.

2.1.2 BIOSTRATIGRAPHIC PROBLEMS

The Ordovician System was first defined by Lapworth (1879), to resolve differences between Sedgwick and Murchison over subdivision of the Welsh Lower Palaeozoic successions (Thackray, 1976). The name of the Caradoc Series derives from Murchison's (1839, p.216-222) Caradoc sandstone exposed northeast and southwest of the hill Caer Caradoc, south Shropshire. No type succession was singled out but the frequent allusions of Murchison to exposures in the River Onny valley near Horderley, SW of Church Stretton, has led to its general acceptance as a stratotype. Bancroft (1933, 1945) proposed subdivision of the Caradoc into 7 stages based on his work in

Shropshire (Fig. 2.2), with the Woolstonian stage being suggested later by Hurst (1979) to replace the upper Longvillian. In spite of this long history of geological and palaeontological examination, there has been little success in erection of a biostratigraphy for the Welsh Ordovician. There are numerous factors which have mitigated against creation of the much needed correlation charts for the Ordovician System.

The Ordovician as a whole, perhaps a function of its tectonic diversity and unrest, was characterized by extreme differentiation of marine faunas (Williams, 1976). The Period contains, for example, 1/3 more families than average for the Early Palaeozoic (Fig. 2.3). To increase problems, a similar proportion of Ordovician taxa consist of sedentary benthic fauna, such as brachiopoda or echinodermata, which are sensitive to environmental control and result in numerous endemic provinces. Although these provinces might assist environmental reconstructions, they hamper stratigraphic correlations. Williams (1974, p.14) has noted the diachronous distribution of many brachiopod stocks between the adjacent Anglo-French and Bohemian brachiopod provinces.

These difficulties are compounded within the Caradoc series. Although the Caradoc is particularly fossiliferous, taxa consist almost entirely of shelly fauna. Considerable emphasis has thus been placed by palaeontologists on the taxonomy and biostratigraphy of brachiopods and trilobites. Proposed biozones (*eg.* Bancroft, 1945), however, have proved extremely provincial, and difficult if not impossible to recognise outside their type areas (Williams, 1963; Whittington *et al*, 1984). The use of shelly fauna for stratigraphic purposes is further complicated by the complex control of sedimentary facies and substrate on benthic palaeocommunities (Pickerill & Brenchley, 1979; Hurst, 1979; Lockley, 1983a). In addition, debris flows and olistrostromes can mix and transport brachiopod fauna, thereby increasing the confusion (*eg.* Beavon, 1963;

Lockley, 1984; Smith, 1987). These depositional processes can also combine fauna of different ages. Finally, because the speciation of dalmanellids and sowerbyellids, two of the most common Caradoc brachiopod genera, is largely based on statistical methods (Williams, 1963), it is difficult to obtain satisfactory comparisons, particularly in North Wales, because of the deformed nature of the specimens and the common imperfect preservation (Diggens & Romano, 1968, p.603; Wright, 1979, Chapt.5).

Although the 'standard' graptolite zones offer potential for more effective correlations, their use in North Wales (Fig. 2.2) is limited due to: 1) the paucity of graptolites within Caradocian sequences, 2) the lack of continuous fossiliferous sections and 3) the unfortunate fact that all Caradocian volcanicity appears to have occurred within 2 or 3 stages of the Caradoc series and at most two broad graptolite zones (*Diplograptus multidens, D. clingani*) which may well overlap (Skevington, 1971 in Howells *et al*, 1978, p.41).

In summary, the endemism of shelly fauna and problems of their relationship with graptolite biozones at most levels, coupled with sparse biostratigraphic data from conodonts and other taxa, have hampered Ordovician biostratigraphic correlations and erection of an acceptable chronostratigraphy (see Bassett, 1984). It is not surprising that, in contrast with Murchison who championed the use of fossil evidence in erecting the "imperialistic" Silurian system , Sedgwick, when faced with the daunting prospect of subdividing the more complex 'Cambrian'(Ordovician), advocated the use of lithostratigraphy in field analysis. His view that palaeontology should be subservient to lithology was forcefully stated in 1831 when he wrote that "[organic remains] can in no instance supersede the necessity of study in detail of the structure and superposition of the great mineral masses covering the surface of the globe".

2.1.3 ACCOMPLISHMENTS OF LITHOSTRATIGRAPHY

The application of lithostratigraphy to North Wales, although more successful than any biostratigraphic approach, has not been spectacular. The inevitable result of a purely lithostratigraphic approach to sequences characterized by remarkably rapid lateral changes in grain size and facies, has been the creation of elaborate local stratigraphic subdivisions, with no obvious correlations between sequences, and the proliferation of stratigraphic nomenclature. The continual redefinition and renaming of lithological units, and periodic elevation of members to formations/groups has increased the confusion (see Fig. 2.4).

Considerable progress has recently been made, however, in unravelling the complex stratigraphy of many volcanic units. As a result of remapping of volcanic sequences in northeast Snowdonia, Howells *et al* (1978,1983) proposed the following, largely lithostratigraphical, subdivision of Ordovician volcanic successions (Fig.2.5, 2.6, 2.7). Two major volcanic groups were defined: the Llewelyn Volcanic Group and the Snowdon Volcanic Group.

The Llewelyn Volcanic Group represents the first major episode of Caradoc volcanism in Snowdonia and varies in thickness from 1200m at Tal y Fan (SH 730 730) to 180m at Gwastadnant (SH 610 580). Volcanic formations defined within the Group (Fig. 2.7, 2.8) are briefly reviewed here as they could provide sources of sediment for the successions studied in this thesis. They include, in approximate stratigraphic order:

1. Braich Tu Du Volcanic Formation: Basic tuffs, rhyolite lava, and acidic ash flow tuffs.

2. Foel Grach Basalt Formation: Thick succession(300m) of basalt lava flows and related breccias.

3. Foel Fras Volcanic Complex: Largely simultaneous with Conwy Rhyolite. Thick accumulation of trachyandesite lavas and ash flows.

4. Conwy Rhyolite Formation: A thick sequence (1000m) of rhyolite lavas, inter-flow breccias.

5. Capel Curig Volcanic Formation: Extends across entire outcrop of the Llewelyn Volcanic Group and represents the first major ash flow volcanism within the Caradoc. The formation has been divided into four volcanic members (Fig. 2.9), the lower two representing single volcanic events. Distribution of the Capel Curig Volcanic Formation, and present correlation of the tuff members is provided in Fig. 2.10. The formation has been the subject of considerable attention (Francis & Howells, 1973; Howells & Leveridge, 1980, etc.); the interpretation of environments of deposition of the tuff members (Chapt. 6), entrance of ash flow tuffs which deposited the first and second volcanic members (Garth and Racks tuffs respectively) into the sea (Chapt. 7.3.2), and the sedimentology of the 4th volcanic member (Chapt. 7.2.4) is a major objective of this thesis.

The Snowdon Volcanic Group comprises a bimodal basalt-rhyolite succession and includes a major caldera forming episode (the Lower Rhyolitic Tuff), superseded by surtseyan and strombolian basaltic volcanism (Reedman *et al*, 1984). The Lower Rhyolitic tuff itself was largely ponded within an asymmetric trapdoor-type volcanotectonic depression (Howells *et al*, 1986; Fig. 2.6). Caldera-fill sequences include a basal welded ash flow sheet, sedimentary slide breccias, non-welded ash flow tuffs and significant thicknesses of reworked tuffs and associated marine sediments.

The intervening Cwm Eigiau Formation was defined (Howells *et al*, 1983) to include all sediments (mudstone, siltstone, sandstone, airfall tuff, local accumulations of basaltic tuff) between the Llewelyn and Snowdon/Crafnant Volcanic Groups. To the south of the Ogwen valley (SH_65_60 to SH_72_59), the formation includes the Pitts Head Tuff(PHT). The PHT was emplaced as a thick (ca. 700m) intracaldera sequence at Llwyd Mawr (SH 508 460; Roberts, 1969) with the lowest outflow tuff extending up to 20 km to the NE to Llyn Ogwen (SH 650 605; Fig. 2.5).

As the Pitts Head Tuff has been recently elevated to formation status (Reedman *et al*, 1987) and can now be identified over a much wider area than originally suspected (see Chapt. 2.2.4.4), it is stratigraphically desirable to split the Cwm Eigiau Formation into two units. The alternative scenario would have the same formation succeeding itself in vertical sequence and violating the Law of Superposition. In this study,

sandstones stratigraphically subjacent and suprajacent to the Pitts Head Tuff formation are informally referred to as the Lower and Upper Cwm Eigiau Formation respectively. At Cwm Eigiau itself (SH 705635), the formations consists largely of poorly exposed cleaved mudstones, recognition of the Pitts Head Tuff horizon is difficult, and palaeontological information is scarce. The type sections (Fig. 2.12) of the Cwm Eigiau Formations are thus best relocated at Llyn Bochlwyd - Cwm Cneifion (SH 655 585 to 646 584) where the division of the previous formation into a lower and upper portion, with contrasting depositional styles, is well displayed.

Further subdivision of the thick sequence of epiclastic volcaniclastics within the Cwm Eigiau Formation was not attempted by the British Geological Survey. In central and southern Snowdonia the sequence underlying the Snowdon Volcanic Group is dominated by sedimentary rocks. The subdivision of these strata has been extremely confused, largely as a result of the inconsistent use of the term 'Glanrafon' (see Howells, Leveridge *et al*, 1983). This same term has been applied to ?Llanvirn and Caradoc sediments below the Pitts Head Tuff on Moel Hebog (Shackleton, 1959), slates to medium sandstones beneath the Snowdon Volcanic Group to the east of the Gwynant valley (Beavon, 1963) and at Dolwyddelan (Williams & Bulman, 1931), and to sequences at Llyn Cowlyd beneath the Crafnant Volcanic Formation (Diggens & Romano, 1968).

In this thesis, a combination of tephrochronology and a depositional systems approach to stratigraphy is used, both to correlate tuff members within the Capel Curig Volcanic Formation, and to partially unravel the ubiquitous Glanrafon sandstones. Correlations using the volcanic units will be considered first.

2.2 TEPHROCHRONOLOGY

2.2.1 INTRODUCTION

Tephra (Thorarinsson, 1954) consists of any pyroclastic accumulations, irrespective of grain size. Although the term 'tephra' was originally defined to apply solely to ash fall material, its usage has since broadened and the term is now synonymous with "pyroclastic material" in general. All tephra layers, irrespective of origin, are emplaced geologically instantaneously; they are commonly deposited over broad regions and thus provide excellent key horizons for wide correlations. These unique time stratigraphic markers have been particularly useful in Quaternary sequences (Westagate & Gorton, 1981). Although the correlation of tephra has been achieved by a wide variety of techniques, it is now recognized that chemical fingerprinting must be an essential component of any scheme.

One of the most significant questions that must be addressed in any use of tephrochronology for stratigraphic purposes is whether or not one would expect to find individual tephra layers to be of uniform composition over broad areas. In airfall ashes, for example, aeolian fractionation resulting from differences in settling velocities of crystals, small pumice particles, glass shards and rock fragments produces downwind increases in SiO_2, K_2O and decreases in Al_2O_3, CaO and TiO_2 within bulk samples (*eg.* Sarna-Wojcicki *et al*, 1981). It is probable that bulk chemical variation resulting from differences in vitric, lithic and crystal contents produced during the eruption, transportation, and deposition of pyroclastic flows will be even more significant (Walker, 1972). The above mentioned physical fractionation necessitates detailed chemical analysis of individual separates (*eg.* glass shards, crystals) to properly characterize volcanic units. Microprobe analysis of individual shards has emerged as one of the most powerful tools in the last decade because of the number of elements which can be analyzed and the great precision obtained by determining shard

to shard variability (see Sarna-Wojcicki *et al*, 1987).

The application of such refined techniques to ancient successions is rarely possible. Devitrification, hydration and recrystallization of glass shards, in connection with diagenesis, hydrothermal alteration, and lower greenschist facies metamorphism both precludes disaggregation of pyroclastic material into mineral separates in addition to obscuring recognition of individual pyroclasts in thin sections. Vapour phase crystallization and sintering of glass shards during welding within many ashflow tuffs in North Wales compounds the problem. This ensures that bulk chemical analysis is the only approach that can be used in tephrochronological studies of such ancient sequences. Although several authors have successfully used bulk analysis of trace elements to correlate ancient bentonites (*eg.* Ordovician, Ohio; Huff, 1983) over considerable distances (300km), it is not certain whether such techniques are generally applicable.

Post-depositional changes can also result in substantial chemical alteration within volcanic rocks. A detailed description of the alteration of Ordovician igneous rocks is given by Kokelaar *et al* (1984a). They stress that petrogenetic and tectonic models of these rocks that rely heavily on whole rock normative analyses or require immobility of the alkali elements (*eg.* Fitton & Hughes, 1970) must be considered as weakly founded. In analysis of basalts and andesites, researchers have found that Ti, Zr, Y, Cr, Hf, Th, and Ta were the least "mobile" elements during chemical alteration. It has thus been concluded (Kokelaar, 1979; Dunkley, 1979; Bevins, 1982) that a similar suite of elements provides the best approximation of primary trace element concentrations in basic volcanics in Wales.

Even less is known about the effects of post-depositional diagenetic and/or metamorphic changes on the mobilization of either major or trace

elements in units of rhyolitic composition, particularly beds of volcanic ash. In an examination of the alteration of acidic tuffs by comparing chemical analyses from whole rock and glass separate samples, some authors (*eg.* Yamamoto *et al*, 1986) have found that TiO_2 normalised plots of various trace (*eg.* Ga, Zr, Co) and even major elements (*eg.* SiO_2, Al_2O_3) were effective in identifying the origin of siliceous sediments. In contrast, Zielinski (1982), in a detailed study of alteration of rhyolitic ash to montmorillonite (Fig. 2.12), indicated mobility of most elements with leaching of U, Si, Na, K, Rb, Cs, and Zr, and concentration of Al, Th. Ta, Hf, Sb, and Ti in an insoluble residuum (5-50% enrichment) and substantial concentration of other elements (*eg.* Mg, Ca, H_2O, Fe, Mn, Sc, Co, Cr, P) in clay through coprecipitation, ion exchange or adsorption. Chondrite normalised plots of rare earth elements were found to be the only useful tool to characterize and identify the silicic ashes; similar conclusions have been reached regarding rhyolitic intrusions in North Wales (A. Kearsley, pers. comm., 1988).

2.2.2 METHOD OF ANALYSIS

Against this background, tephrochronology was attempted with caution in North Wales in the hope of correlating ash-flow tuffs with their distal equivalents and with possible co-ignmibrite air fall ashes. Specific aims included:

1. identification of the distal portion, if any, of the Pitts Head Tuff.

2. verification of the concurrent eruption of the 2nd and 4th members of the Capel Curig Volcanic Formation as suggested by the occurrence of vitric ash flow tuffs macroscopically reminiscent of the 2nd volcanic member within the 4th member at Fynnon Lloer (Howells *et al*, 1980; this study, Fig. 2.10b).

3. correlation of Glanrafon sandstones at Moel Hebog and Yr Arddu with

Llewelyn Volcanic Group/ Cwm Eigiau Formation in NE Snowdonia.

Over 200 samples, weighing approximately 2 kg each, were collected from rhyolitic ash flow tuffs and tuffites. A total of 113 representative samples were selected for geochemical analysis (Appendix 1.1), with care taken to choose specimens with the least obvious alteration or fracturing. Powders from an additional 9 samples of the Pitts Head Tuff collected by the British Geological Survey (BGS) were also analyzed for trace elements to compare results obtained by the Nottingham/Oxford labs. Correlation coefficients and linear regression of individual elements from this comparison is presented in Appendix 1.2. Full details of both sample preparation and preparation of pellets and beads for XRF are given in Appendix 1.3. A summary table of the location and/or identity of various tuffs examined is shown in Fig. 2.13, with positions of field localities indicated on the map in Fig. 2.14.

All 124 powders available were analyzed for 13 standard trace elements (Rb Sr Y Zr Nb Pb Zn Cu Ni Co Cr V Ba) using XRF; only 61 of these samples were analyzed by the same technique for major elements. The potential ability of major elements to characterise and stratigraphically correlate volcanic units will be discussed first with the usefulness of trace elements to achieve these purposes considered later (Chapt. 2.2.4). All data were treated by univariate and multivariate analysis of variance, and stepwise discriminant analysis using subprograms (MANOVA, ONEWAY, DISCRIMINANT) within the Statistical Package for the Social Sciences (SPSS-X).

2.2.3 STRATIGRAPHIC CORRELATIONS USING MAJOR ELEMENTS

Raw geochemical data for all XRF analyses of major elements is presented in Appendix 1.4. Normalised oxide percentages (Appendix 1.6) were used to rectify variation induced by loss of flux during melting (see Appendix 1.5); a summary of this data grouped by tuff and/or sampling

locality is presented in Figure 2.15. Any attempt to use major element chemistry for stratigraphic purposes, must consider both the extent of major element variation within single flows due to fractionation of crystal and glass phases during deposition of the pyroclastic flow and/or different diagenetic histories, as well as determining whether major element compositions are chemically unique for each ash flow tuff.

Insufficient samples from a single flow were analyzed for major elements to properly assess the extent of variation resulting from flow fractionation or diagenesis. A crude idea of this chemical change, however, can be obtained by comparing the major element composition of the 3 known tuffs with all other tuffites, the distal representatives of the ash flow tuffs (Fig. 2.16, 2.17). Most of the ash flow tuffs/tuffites would be classified as high silica rhyolites (Hildreth, 1981) containing greater than 75% silica. Many of the distal ash flows or tuffites, however, contain silica percentages greater than 77.4%. It has been suggested that no unaltered rhyolite should contain greater than this amount of SiO_2, when reasonable magmatic values are assumed for halogens and unanalyzed minor elements (Hildreth, 1981). The high percentages of SiO_2 in this case within "tuffites" is probably a result of the greater vitric component found in these rocks compounded by later silicification in a marine environment. The higher percentages of MgO, CaO and Na2O, are likely to be a function of marine alteration. The significant reduction in K_2O within "tuffites" probably arises from the lack of sanidine or orthoclase, which is a common crystal phase within "proximal" ash flow tuffs. The significant variation suggested by these major element comparisons would, of course, hamper any stratigraphic correlations.

A comparison of the major element composition of the 3 known tuffs (Fig.2.18), similarly suggests that use of major elements for stratigraphic correlations is pointless. Multivariate analysis of variance is a useful

statistical technique to compare two or more sets of data with many variables such as above. The test statistic used to compare vector means is Hotellings T^2 which can be expressed as follows, for two samples:

$$T^2 = n\,[X-u]' * [s^2]^{-1} * [X-u]$$

where X = vector mean for a sample
u = vector mean for other sample
$[s^2]^{-1}$ = inverse of variance-covariance matrix.

The critical values of T^2 can be determined by the relationship

$$F = \frac{n1 + n2 - m - 1}{(n1 + n2 - 2)m} \; T^2$$

where n = number of samples and m = number of variables, allowing comparisons to be made with more conventional univariate F tables. The univariate test statistic for comparison of group means, F, can be expressed as:

$$F = \frac{MS_A}{MS_B}$$

where MS_A is the variance among different groups and MS_B is the variance between different groups.

Although a multivariate comparison of vector means reveals that the major element composition of the Garth, Racks and Pitts Head Tuffs are statistically different with significance levels less than .0005 (Fig. 2.19a), an examination of individual elements (Fig. 2.19b) shows that K_2O, Na_2O, P_2O_5, and TiO_2 are the only major elements with any univariate statistical variation between the three tuffs. An additional comparison of these elements (Fig. 2.20) using the Tukey-B procedure (as part of ONEWAY subprogram), indicates that the most chemically unique element P_2O_5 (highest F ratio; Fig. 2.19b), is only able to discriminate between the Garth or Racks Tuff and the Pitts Head Tuff but not the individual members of the Capel Curig Volcanic Formation; TiO_2, commonly recognized as an immobile element, only displays statistically significant differences in mean concentration between the Racks Tuff and Pitts Head Tuff. At the

other end of the spectrum, none of the ash flow tuffs contain statistically different amounts of SiO_2, Al_2O_3, Cr_2O_3, Fe_2O_3, MnO, MgO or Na_2O. It thus appears that major elements offer little hope in either discriminating between ash flow tuffs, or stratigraphically correlating ash flow tuffs in North Wales. Faced with the failure of major elements to achieve these aims, one is forced to examine the possibility of using the trace element composition.

2.2.4 STRATIGRAPHIC CORRELATIONS USING TRACE ELEMENTS

2.2.4.1 INTRODUCTION

Raw geochemical data from trace elements are presented in Appendix 1.7, with a summary by tuff and/or sampling locality shown in Fig. 2.21. Replicate analyses of sample G075 (Fig. 2.22) reveal that the XRF analysis appears to be the most precise for trace elements with the highest concentrations, with Zr, Rb, Sr, Nb and Y being particularly consistent between portions.

Sixty-two (62) of the total 124 samples analyzed for trace elements represent "known" ash flow tuffs. The "parent groups" which were analyzed and the number of representative samples, are as follows:

Capel Curig Volcanic Formation:	1st member (Garth tuff)	10 samples
	2nd member (Racks tuff)	8 samples
	3rd member	2 samples
	4th member (Dyffryn Mymbyr tuff)	6 samples
Pitts Head Tuff Formation:		36 samples

Any attempt to utilise whole rock analysis of trace elements from these pyroclastic rocks to classify the remaining unknown entities into these various parent groups and thereby establish a regional stratigraphy must again consider both 1) the extent of lateral variation in chemistry within individual ash flow tuffs and whether distal equivalents of pyroclastic units would retain their chemical signature, AND 2) whether

the known ash flow tuffs are chemically unique and can be distinguished from each other.

2.2.4.2 TRACE ELEMENT VARIATION WITHIN AN ASH FLOW TUFF

Samples were collected from the Pitts Head Tuff at 4 localities (Fig. 2.14),2.25 to assess lateral variation within ash flow chemistry. The localities, distances from assumed source at Llwyd Mawr (Reedman *et al*, 1987), depositional environment (Orton, 1987; Chapt. 4.3.7 & 6.2.5), and number of analyses available (using the 9 BGS powders), are tabulated below:

REG	LOCALITY	DISTANCE FROM LLWYD MAWR	DEPOSITIONAL ENVIRONMENT	NUMBER OF SAMPLES
1	Cwm y Straddlyn	5.5km	alluvial fan	2
2	Pont-y-Gromlech	17km	upper shoreface	11
3	Gribin Ridge	19km	lower shoreface	13
4	Ogwen Cottage	20km	lower shoreface	10

Means and standard deviations of trace element content of these sampling regions are listed in Fig. 2.23.

A multivariate statistical comparison of the Pitts Head Tuff using all 13 elements, shows that the bulk trace element composition of the Pitts Head Tuff is statistically different between the 4 regions of study, with high significant levels of <.0005 (Fig. 2.24a). This can be seen graphically by a plot of the chemical composition for individual elements normalized against the bulk mean for the Pitts Head Tuff (Fig. 2.25 a,b). Locality 5, at Pen-yr-Helgi Du, represents an ash flow tuff chemically correlated with the Pitts Head Tuff (Chapt. 2.2.4.4). Several elements show significant lateral variation within the ash flow tuff with Sr, Co displaying systematic increases away from source and Rb decreasing in this direction. Differences within trace element composition of Cwm Ystraddlyn in relation to the Pitts Head Tuff mean are likely to be in part merely a function of the small number of samples(2) analyzed. The "Eu" anomaly

style of

within the diagram" at Bochlwyd ridge with marked increases in Rb, Pb, and Sr and decreases in Cr is a chemical manifestation of ingestion of significant quantities of water within the Pitts Head Tuff related to concurrent intrusion of several thick (1-10m) sandstone dykes into the ash flow at this locality (Chapt. 7.2.5). It can be seen that Nb, Y, and Zr display the least lateral variation within the Pitts Head Tuff, particularly in the distal portion from Pont-y-Gromlech to Ogwen Cottage, and are among the most reliable elements to use for any stratigraphic correlation.

These suggestions are supported by a univariate comparison between regions (Fig. 2.24b). Elements with low F ratios, and little lateral variation within the Pitts Head Tuff, include, in order of increasing variance, include Zr, Nb, Cr, Co and Cu. Most of the variance within Yttrium results from the anomalously low values at Cwm Ystraddlyn (26.7ppm compared with 49.16ppm PHT mean); low F ratios (.23489; F critical = 5.36 at 99% confidence level) and correspondingly high significance levels (.792) are obtained if one excludes this locality from comparisons. If one again treats data to a multivariate comparison of means (Fig. 2.24c), and only uses the elements Zr, Nb, Cr, Cu, and Y, one finds that there is NO statistical difference laterally in the trace element composition of the Pitts Head Tuff. It is encouraging that elements which display the least lateral variation within the Pitts Head Tuff and are also the most "immobile" (*eg.* Pearce & Cann, 1973; Winchester & Floyd, 1977), and least susceptible to leaching or concentration resulting from different diagenetic histories (subaerial, shallow marine environments etc.). It thus appear that restriction of multivariate comparisons between ash flow tuffs to these comparatively immobile elements should ensure that statistical differences in vector means between samples are "real" and not a function of either diagenetic or pyroclastic "fractionation".

2.2.4.3 CHEMICAL VARIATION BETWEEN ASH FLOW TUFFS

Means and standard deviations for the 5 "known" ash flow tuffs are listed in Fig. 2.26. A multivariate comparison using all thirteen trace elements indicates that the 5 ash flow tuffs have statistically different trace element compositions (Fig. 2.27a); a comparison using individual elements reveals that Zr and Y are the most reliable elements (high F ratios) for distinguishing between ash flow tuffs(Fig. 2.27b) with Zr alone able to separate all tuff combinations except the Garth and Racks Tuff, and the 3rd and 4th members of the Capel Curig Volcanic Formation (Fig. 2.27c). A triangular plot (Fig. 2.28) of the relative proportions of the three most useful elements (Zr, Y, Nb) displays the distinct chemical trace element concentrations of the "known" ash flow tuffs.

Stepwise discriminant analysis using the Wilks lambda method was used to establish the most useful elements for discriminating between the "known" ash flow tuffs. Discriminant analysis is a useful multivariate technique that calculates the effects of independent variables (*ie.* elements) on a nominal dependant variable (*eg.* tuff classification). In the stepwise mode, elements are entered into the discriminant functions in order of the explanatory power using Wilks lambda. This statistic is a ratio of within group variance to the total sum of squares; it will approach zero if variance within tuffs is small compared to total variance, and most variance is due to chemical differences between ash flow tuffs. It can be seen (Fig. 2.29) that the most successful elements for distinguishing between the 5 ash flow tuffs are, in decreasing order of discriminatory ability, Zr, Y, Nb, Cr, and Cu. Eigenvalues, associated canonical correlations and cumulative percentages of discrimination accounted for in each successive function calculation are shown in Fig. 2.30, with unstandardized discriminant function coefficients tabulated in Fig. 2.31.

Verification of the high degree of strength inherent in such a discriminant model is provided by classification of the known tuffs. Discriminant scores, and classifications of the 62 individual "known" samples are listed in appendix 1.8. The correct identification of tuff samples was predicted 98.4% of the time, with only one sample, G078, being misclassified. Sample G078 represents an extremely fine-grained vitric tuff of cryptocrystalline quartz with anomalously low Zr(75.3 ppm) and Nb(12.0 ppm) values and high Y (83.3 ppm) values, collected from the extreme top of the Pitts Head Tuff at Pont-y-Gromlech; its significance will be discussed later (Chapt. 2.2.4.4 & 7.3.2.1).

2.2.4.4 CLASSIFICATION OF UNKNOWN SAMPLES

A classification of all unknown samples (see Fig. 2.28 as well) into one of the five known categories is presented in Fig. 2.32. Although many of the suggested correlations are spurious, with samples being "forced" into physically impossible classifications, the groupings provide useful guidelines for potential stratigraphic correlations and further statistical treatment. The validity of suggested correlations can be assessed through use of multivariate comparison of two sample vector means (see Chapt. 2.2.4.2 for meaning of statistical tests); results of such analysis for most possible tuff pairs are tabulated in Fig. 2.33, The ramifications of individual cases are now discussed in detail; the greatest faith was placed in correlations suggested or refuted using the five immobile elements.

1. There is no statistical difference using Zr, Y, Nb, Cr, and Cu between the Garth Tuff and the ash flow tuff at Rolwyd (tuff 101) supporting the earlier suggestions (Francis & Howells, 1973, p.629) for this correlation; the composition of the Rolwyd tuff is statistically different from that of the Racks tuff at a 99% confidence level.

2. Both the Lledr Valley tuff (103) and sample G095 (tuff 105) from the lowest ash flow of the Capel Curig Volcanic Formation (CCVF) on Carnedd

Dafydd (Fig. 2.10b, 2.14) show no statistical difference in trace element composition from the Garth tuff and are chemically distinct from all other members of the CCVF.

3. A thin airfall tuff (1 sample - tuff 104) within deltaic sequences on Moel Ddu is chemically distinct from any member of the CCVF, including the 4th volcanic member with which it was grouped (Fig. 2.32).

4. The second Lledr Valley tuff (sample 118: tuff 201) cannot be assigned to either the 1st or 2nd volcanic members of the CCVF with high confidence levels, although using Zr, Y, Nb, Cr and Cu it shows more chemical similarity (lower F ratio; higher significance level) with the 1st member(Garth Tuff) than the Racks tuff. Although the correlation is not significant, it is suggested that the second Lledr Valley Tuff is a faulted repetition of the Garth Tuff, rather than a distal representative of the Racks tuff (Howells & Leveridge, 1980, p.3).

5. The vitric rich ash flow tuffs at Ffynnon Lloer (tuffs 202/203) within unquestionable 4th volcanic member of the CCVF tuffs, are chemically distinct at a 99% confidence level (using Zr, Y, Nb, Cr and Cu) from the 2nd volcanic member of the CCVF (and the 1st member as well) with which they were previously presumed to correlate (Howells & Leveridge, 1980, Fig.2). One can not therefore explain the deflection of the 2nd member of the CCVF toward the southeast (*eg.* Howells & Leveridge, 1980, p.14) by the developing pile of the 4th member of the CCVF, but must seek other mechanisms (Chapt. 6.2.2). The Ffynnon Lloer ash flow tuffs (202/203) show the greatest chemical affinity with the 3rd member of the CCVF, confirming that the last two members of this volcanic formation were partially coeval in their eruption (see Chapt. 7.2.4 & 7.4).

6. The Pitts Head Tuff (PHT) is chemically distinct from all possible tuff/tuffite correlatives in section F (Craig yr Ysfa to Pen Llithrig yr Wrach), section H (Llyn Colwyd), and section G (Capel Hill) **except** tuff F2

(number 502), a vitric rich primary ash flow tuff (depicted on the 1:25,000 Passes of Llanberis and Nant Ffrancon Map), with which it shows NO statistical difference in composition using either all 13 trace elements or the 5 "immobile" elements (Fig. 2.34). This correlation of the PHT an additional 5 kilometres into the basin (Fig. 2.14, S63, has important implications regarding regional stratigraphy (Chapt. 2.4), tectonic evolution of the basin (Chapt. 6), and entry of ash flow tuffs into the sea (Chapt. 7.3). It also implies that most of the sequence to the east of the A5 road is stratigraphically younger than the PHT; if placement of the Soudleyan-Longvillian boundary (Diggens & Romano, 1968) is correct (high within sequence, *ie.* near tuff Cowlyd 2(512), Llyn Cowlyd; tuff G6 (526), Capel Hill), then the PHT was erupted well within the Soudleyan. The Soudleyan-Longvillian boundary is placed within the type section of the Cwm Eigiau Formation (Fig. 2.11), using brachiopod determinations by S.P. Tunnicliff (BGS Open File Report[1], PD 80/27), who notes a changes from Soudleyan fauna (*D. berwynensis, Plaesiomys multifida* etc.) towards a Longvillian fauna (*Sowerbyella cf. soudleyensis*) within sediments at Cwm Cneifion. The exact positioning of the stage boundary between BGS fossil localities 4 (op. cit., item 10; SH 6458 5832) and 10 (op cit, item 9; SH 6463 5848), a stratigraphic distance of *ca.* 40 m, is uncertain.

7. The intriguing positioning of Tuff 102 in the midst of the PHT field on the triangular Zr-Y-Nb plot is thought to be real. Although a multivariate comparison of the absolute amounts of trace elements (Fig. 2.36a) indicates that the two most reliable elements (Zr, Nb) are statistically different (high F ratios), a univariate comparison using the ratios of trace element compositions over Zr reveals no statistical difference in Y/Zr or Nb/Zr ratios (Fig. 2.36b). A multivariate comparison

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[1]All BGS Open file reports used in thesis are reproduced for those interested in Appendix 2.

using only Y/Zr and Nb/Zr ratios confirms a closer affinity with the PHT than the 3rd volcanic member of the CCVF which is suggested (Fig. 2.33) using absolute trace element concentrations. In fact, Tuff 102 has identical absolute concentrations of Nb, Cr, V and Ba to the anomalous airfall sample (GO78) from the top of the PHT at Pont-y-Gromlech (Fig. 2.36d); it is suggested both these unusually fine-grained vitric tuffs are a by-product of explosions generated by entry of the PHT into the sea (see Walker, 1979).

8. There is little statistical difference in chemical composition (Fig. 2.33) between pairs of tuffs stratigraphically younger than the Pitts Head Tuff among sections F (tuffs 503-505) at Bwlch Trimarchog, G (tuffs 522-526) at Capel Hill or H (tuffs 511, 512) at Llyn Cowlyd. Most tuffs probably represent individual depositional lobes. It is suspected that use of additional samples or more importantly more "immobile" trace elements would yield a few correlations. A stratigraphy for these successions is later suggested (Chapt. 5.4.1.5) by correlation of depositional systems and consideration of faunal evidence.

2.2.5 CRITIQUE OF TEPHROCHRONOLOGY

The use of tephrochronology for stratigraphic purposes in North Wales was only partially successful. The chemical variation between vitric rich tuffites and ash flow tuffs within the Cwm Eigiau Formation in eastern Snowdonia was insufficient to allow discrimination and correlation between these volcanic units.

As mentioned earlier, a more substantial source of error that must be addressed was chemical variation produced by fractionation of vitric/crystal components during emplacement and deposition of the pyroclastic flow. The difficulties in correlating the extremely fine-grained crystal-depleted "phreatomagmatic" ash overlying the Pitts Head Tuff at Pont-y-Gromlech with the ash flow itself have already been noted.

The significance of chemical fractionation is further underlined by the fact that both tuff 530, logged as reworked Pitts Head Tuff (section D2, 8m), and the unquestionable distal portion of the Pitts Head Tuff at Llyn Cwm-y-fynnon (tuff 501), are chemically distinct at a 99% confidence level (using all 13 or only the 5 immobile trace elements) from the Pitts Head Tuff proper (Fig. 2.40A,B). These differences in chemical signature between the Pitts Head Tuff and its distal or reworked equivalents can be attributed to the higher crystal content of the latter, which would account for the high Zr and Nb concentrations. The physical changes that can be experienced upon entry of an ash-flow tuff into the sea necessitate great caution in correlating vitric-rich subaerial ash-flow tuffs with either crystal-rich or crystal-depleted subaqueous equivalents; comparative microprobe analyses of individual feldspar crystals or microscopic removal of crystals from crushed aggregates must be attempted before reliable correlations can be made.

Despite the above limitations, tephrochronology still offers the best hope of future erection of a chronostratigraphy in North Wales. In this pilot study, successful discrimination between and correlation of primary ash flow tuffs was achieved, in essence, using only three trace elements (Zr, Y, Nb). Most researchers attempting geochemical correlations (*eg.* Huff, 1983) use many additional immobile trace and rare earth elements, including Th, Ta, La, Hf, and Dy, usually determined by Induced Neutron Activation (INNA) or atomic absorption. The inclusion of these elements within comparisons of chemical signatures for ash flow tuffs/tuffites in North Wales would result in an exponential increase of the discriminatory power of geochemistry. The implications for our understanding of the stratigraphic and tectonic evolution of the Welsh basin are obvious: it is unfortunately beyond the scope of this thesis to even begin to address such a problem.

2.3. CORRELATION OF DEPOSITIONAL SYSTEMS

2.3.1 PHILOSOPHY

Traditionally, stratigraphic successions have been regarded as a "layer-cake" with sediments viewed as geometrically uniform blankets bounded by sharp horizontal lines on all correlation tables. Beginning in about 1960, however, fundamental changes in sedimentary geology, including use of the facies model and the depositional systems concepts, seismic stratigraphy, refinements to chronostratigraphy and provision of a plate tectonic framework (see Miall, 1984, p.3-5) have necessitated development of a completely different genetic approach to the subject of stratigraphy. The concept of depositional episodes and the depositional system approach to basin analysis was initiated largely in the Gulf coast region, USA, using conventional subsurface data (*eg.* Brown & Fisher, 1977; Galloway, 1981) as a means of analyzing and interpreting the immense thicknesses of oil-bearing sediment. Its application to other sedimentary basins has been encouraged by recognition of analogous lensoid sedimentary packages using modern seismic-stratigraphic techniques (*eg.* Mitchum *et al*, 1977). A depositional system is a 3-dimensional assemblege of lithofacies genetically linked by active (modern) or inferred (ancient) processes and environments. The concept is essentially an extension of Walther's law which states that in a conformable succession the only facies that can occur together in vertical succession are those which can occur side by side in nature . In North Wales, a prograding fan delta is a good illustration of this containing alluvial plain, shoreline, delta front and prodelta facies which co-existed side by side. Depositional systems are bounded either by unconformities or by facies transitions into adjacent genetically unrelated systems.

A similar genetic approach to stratigraphy was recently adopted by Plint *et al* (1986) for the Cardium (Turonian, Upper Cretaceous) in the

subsurface of Alberta. They recognized numerous intertonguing marine and nonmarine coarsening upwards sequences which pinch and swell across the basin. Correlation of the bounding erosion surfaces and related suprajacent transgressive sequences allowed later analysis of the importance of sea level fluctuations during deposition (Plint & Walker, 1987; Bergman & Walker, 1986). This use of an "event" stratigraphy was met by much scepticism in the geological community (Hayes & Smith, 1987; Rine *et al*, 1987) who preferred to interpret units in a purely lithostratigraphic fashion. The use of the depositional system concept, however, enables stratigraphic units to be established on a basin wide scale, predictions to be made about large masses of sediment, and interpretations of rocks in terms of broad regional environmental and palaeogeographic reconstructions.

2.3.2 APPLICATION TO NORTH WALES

Correlation of depositional sequences (chapter 4 to 6) into depositional systems, combined with local correlation of more minor episodes of volcanic activity such as air fall tuffs or tuffites, has allowed subdivision of the thick sequences of volcaniclastic sandstone intercalated with the major ash flow tuffs. Individual units are named according to their geographical location with stratigraphic rank based on thickness and areal extent. Thick (usually greater than 100m) deltaic depositional systems, for example, are given member status, whereas minor air fall tuffs or stratigraphic units which can only be correlated a short distance are given the status of a bed. The purpose of this exercise is not to introduce a plethora of additional stratigraphic terms. Units are only named if they can be correlated between regions, or contain sedimentary information worth describing. The creation of this nomenclature is necessary to provide a stratigraphic framework in which to "dismember" the basin's evolution; it is very difficult to compare, contrast and write

about unnamed parts of the stratigraphy!

Only a generalized framework is presented in this chapter with more localised correlations of sequences for individual depositional systems introduced later (Chapt. 4-6). The stratigraphy is included with an accompanying map as a small desk-top enclosure such that it is readily accessible to use in conjunction with more detailed figures in the following chapters.

2.3.2.1. STRATIGRAPHIC SUBDIVISION OF THE LLEWELYN VOLCANIC GROUP

Epiclastic sequences within the Llewelyn Volcanic Group (Enclosure 1) can be subdividing into two formations depending on their position relative to the Gwern Gof Tuff Formation. The Ddeugwm Formation extends from the stratigraphically lowest unit of primary volcanicity in the Tryfan Anticline to the base of the Gwern Gof Tuff Formation and consists of mudstone to medium sandstone. The Tryfan Formation, which varies in composition from siltstone to conglomerate, extends from the top of the Gwern Gof tuff to the base of the Capel Curig Volcanic Formation (CCVF) and can be traced laterally from Tal y Fan to Llynnau Mymbyr, a distance of 15 kilometres. More refined subdivision of these two formations, as well as the CCVF into members is also shown on the enclosure; the sedimentology of the various depositional systems is discusssed fully in Chapter 4 to 6.

2.3.2.2 STRATIGRAPHIC CORRELATION AND SUBDIVISION OF THE CWM EIGIAU FORMATION

One of the most challenging stratigraphic tasks in North Wales is correlation of the thick sequences of Caradoc "Glanrafon" sandstones and siltstones in SW Snowdonia at Moel Hebog (SH 565 470) and Moel Ddu (SH 579 442) with sediments near Blaen Nantmor (SH 640 485) and successions within the Llewelyn Volcanic Group and Cwm Eigiau Formation in NE Snowdonia. The removal of much critical sedimentary information by formation of the Lower Rhyolitic Tuff caldera in central Snowdonia means that correlation of

individual depositional systems across this intervening area is in part speculative. It is also appears (Chapt. 3.4.2, 6.2.4) that SW Snowdonia represents a discrete block-faulted sub-basin of the Welsh marginal basin, genetically unrelated to basins in NE Snowdonia, for large parts of the Caradoc period.

Correlations can be proposed, however, by; A) taking into account regional tectonic changes related to magma diapirism (Chapt. 6.3), B) using a circuitous stratigraphic route (Tryfan anticline - Capel Curig Anticline - Dolwyddelan syncline - Yr Arddu syncline - Moel Ddu- Moel Hebog) to "get around" the Snowdon problem (Enclosure 1) and C) ensuring that available palaeontological data (Wright, 1979; BGS, Appendix 2) does not conflict. A combination of these three techniques enabled stratigraphic subdivision of the sediments in SW Snowdonia according to terminologies previously erected in NE Snowdonia. Important correlations to note here are: 1) that the Gorllwyn Grits and Gorllwyn slates of Shackleton (1959) on Moel Hebog and Moel Ddu (Fig. 2.4) are the stratigraphic equivalents of the Lower Cwm Eigiau Formation (see Chapt. 6.2.4) and 2) that the Costonian to Soudleyan Prenteg grits and slates are tenatively correlated with the Llewelyn Volcanic Group in NE Snowdonia.

Chapt 3. **SANDSTONE PETROGRAPHY: THE VOLCANIC CONTRIBUTION**

> "I cannot greatly honour the minuteness of details, so long as there is no hint to explain the relation between things and thoughts; no ray upon the metaphysics"
>
> *Emerson, Nature, 1836.*

3.1 INTRODUCTION

3.1.1 TERMINOLOGY

Volcanic eruptions are a major source of detritus in many tectonic settings, usually through the production of abundant pyroclastic debris. The nature of this volcanic contribution to the sedimentary record, however, is extremely diversified. Pyroclastic fragments can vary in chemical composition from basalt to rhyolite, vary in grain size from fine ash to blocks or bombs (Fig. 3.1), and consist of either crystal, lithic or vitric components (Fig. 3.2). Fuller discussions of volcaniclastic petrography and classifications are provided by Pettijohn *et al* (1973, Chapt 7), Wright *et al* (1980) and Schmid (1981).

Sands rich in volcanic debris are of two dominant types according to the method of clast formation. Pyroclastic material is generated by explosive volcanicity, and is usually contributed directly to the sedimentary record as primary pyroclastic deposits. In contrast, epiclastic volcaniclastics result from erosion and weathering of older volcanic rocks, including lithified tuffs, and do not usually reflect contemporaneous volcanicity. A "grey area" within this classification concerns reworking of unconsolidated pyroclastic debris; many authors (particularly Fisher, 1961, 1966; Fisher and Schmincke, 1984, p.89) regard recycled pyroclastic material, redeposited by wind or water as secondary pyroclastics. In this study, volcaniclastics are only termed pyroclastic where clast shapes assumed during volcanic eruption have not been altered beyond recognition by later reworking (*cf.* Schmid, 1981).

Characteristic petrographic features indicative of volcanism within a basin include glass shards, pumice and scoria, bipyramidal β-quartz, and euhedral zoned feldspars. Their recognition, however, is hindered by diagenetic alteration and destruction of diagnostic mineralogy and texture. Volcaniclastic sands are more susceptible to diagenesis than any other type of sand because of the chemical instability and reactivity of their framework grains(Pettijohn *et al*, 1973, p.267) with chalcedony, opal, zeolite and clay minerals being typical early cements and alteration by-products (*eg.* Surdam & Boles, 1979; Davies *et al*,1979). In modern volcaniclastic sediments in Guatemala, such diagenesis has resulted in eradication of many sedimentary structures within 3-5 years of exposure and total occlusion of any original permeability within 2000 years of deposition (Vessel & Davies, 1981; Davies *et al*, 1979). The alteration of lithic fragments and feldspars (Cummins, 1962; Brenchley, 1969) within Ordovician volcaniclastic sandstones led to their classification as greywackes and involvement in discussions on the origin of this type of sandstone (Chapt. 1.2.2). Resultant diagenetic matrix (largely epimatrix) comprising on average 20% of sandstones point-counted in this study (Chapt. 3.3.4). Although later lower greenschist facies metamorphism has altered early diagenetic products to either quartz-sericite aggregates, or chlorite, the amount and composition of any matrix present (Chapt. 3.3.4) were still useful in understanding the original detrital composition of the sandstones.

A further complication is that many volcaniclastic deposits consist of mixtures of both pyroclastic and epiclastic material (Fig. 3.2a). Schmid (1981) referred collectively to mixtures of pyroclastic and epiclastic material as "tuffites" (between 25 and 75% pyroclastics by volume) and called deposits of this nature tuffaceous sandstones or siltstones (Fig.3.2a). The recognition of pyroclastic debris where such material is

"diluted by and mingled with sands of other origins is not easy" (Pettijohn *et al*, 1973, p.234). Resolving this problem, however, is far from an exercise in semantics (Fisher & Schmincke, 1984, p.89), but is crucial in determining the nature of the volcanic contribution to sedimentation. A solution can only be achieved through a combination of detailed facies analyses of rock associations (Chapters 4,5) and use of petrographic data on the physical aspect, composition and relative percentages of various clast types within sediments (this chapter).

3.1.2 AIMS

The volcanic contribution to epiclastic sedimentation in North Wales has long been recognized. Ramsay (1866, p.162) for example, noted that much ashy material was "recomposed, having first fallen on land around one or more of the volcanoes from whence mixed with ordinary sediments, it was spread abroad in the manner of other strata by the effects of marine denudation". Most volcanic rocks in North Wales represent pyroclastic deposits (Chapt. 1.2.3) and lavas of either rhyolitic or basalt composition were rare. In addition, preserved primary or secondary pyroclastics were either contained within penecontemporaneous volcanotectonic depressions (*eg.* Pitts Head Tuff at Llwyd Mawr, Reedman *et al*, 1987; Lower Rhyolitic Tuff, Howells *et al*, 1986), or consist of welded ash flow tuffs (*eg.* 1st and 2nd volcanic members of CCVF, Howells *et al*, 1979; Pitts Head Tuff outflow sheets).

In this light, one must query whether the thick sequences of assumed epiclastic material intercalated with the obvious units of primary or secondary pyroclastics, actually represent redeposition of loose unlithified volcanic ash which themselves also record episodes of volcanicity. In most convergent margin volcanic arcs (*eg.* Mathisen & Vondra, 1983; Kuenzi *et al*, 1979), for example, episodes of alluvial fan or deltaic progradation occur during episodes of volcanicity. The alternative

would have periods of epiclastic sedimentation merely reflecting tectonic changes within the basin with associated reworking of both arkosic basement (*ie.* granite, gneiss) and/or previously deposited consolidated volcaniclastics.

In this chapter, an attempt is made to evaluate the nature (*ie.* direct or indirect) and extent of the volcanic contribution to the petrography of clastic sediments intercalated within known pyroclastic deposits. Over 300 thin sections were examined petrographically, from both primary and secondary pyroclastic deposits, as well as volcaniclastic sandstones and siltstones, with 92 samples, representing all parts of the stratigraphy, selected for analysis by point-counting. Both standard optical and back-scatter SEM microscopy were used.

The petrography of known primary and secondary pyroclastic units, of both rhyolitic and basaltic composition, will be briefly examined first to observe the composition, texture, and diagenetic alteration of pyroclastic material produced within the basin. The composition of ?epiclastic sandstones will then be compared (Chapt. 3.3, 3.4) both to comment on the relative importance of volcanic and "basement" sources, and to document the regional and stratigraphic variation in the type and amount of volcanic input. Sandstone petrography will later be used to assess the validity of the recent plethora of ternary discriminant diagrams (*eg.* Dickinson & Suczek, 1979) purporting to provide information about the plate tectonic setting of depositional basins from sandstone petrography.

3.2 PYROCLASTIC PETROGRAPHY

Primary and secondary pyroclastic units are extremely varied in the grain-size, composition and texture of their constituent fragments, as well as the amount of epiclastic material incorporated before deposition. General microscopic features are described below, with characteristics of

individual ash flow tuffs discussed later (Chapt. 7) with reference to their depositional environment.

3.2.1 PRIMARY PYROCLASTICS

Rhyolitic ash flow tuffs are the dominant expression of primary pyroclastic volcanicity. Typically, the tuffs are pale green on fresh surfaces and bleached when weathered (Fig. 3.2b). They are often welded. In the Pitts Head Tuff a planar eutaxitic foliation fabric exhibited by flattened chloritic fiamme is easily discernible on weathered surfaces (Fig. 3.3), whereas in most members of the Capel Curig Volcanic Formation one has to resort to thin sections to identify welded shard fabrics (Fig. 3.4, 3.5). Non-welded vitroclastic fabrics are often present within the top portion of ash flow tuffs; for example, within the upper one-third of the Garth tuff in the Capel Curig anticline (SH 7062 5769; Fig.3.6, 3.7a,b).

The primary ash-flow tuffs are shard-rich with crystals of feldspar (albite-oligoclase or sanidine) up to 3mm long or smaller (1-2mm maximum) quartz crystals comprising (except in basal concentration zones) less than 5% of the total volume. The shards, are extremely variable in shape and size, with angular to slender cuspate forms recognized (Fig. 3.6, 3.7a). Lithic fragments, mostly siliceous siltstone or sandstone fragments less than 30cm in diameter, are rare or absent. They are usually derived locally from subjacent strata, and predominantly found where changes within underlying topography caused flow transformations within the ash flow tuffs (Chapt 7.2.5). All ash flow tuffs are completely devitrified and recrystallized, with shards replaced by a microcrystalline aggregate of quartz, sericite, and feldspar, with minor chlorite. Alteration commonly results in total obliteration of shard fabric (Fig. 3.8, 3.9); in the Pitts Head Tuff, perlitic fracture of originally glassy matrix is usually the only microscopic texture visible. Mobility of silica associated with devitrification has resulted in the common formation of siliceous nodules

within ash flow tuffs (Fig 3.10). The characteristics of nodules vary considerably; some represent infill of vesicles, while others accentuate a welding fabric concordant with, or partially deflected by their outer surfaces (Fig. 3.11). The above relationships indicate an early formation of nodules, before or during compactional welding, in many cases. In thin section, nodules comprise a quartz mosaic with depletion of silica obvious in zones between nodules (Fig. 3.11).

Most non-epiclastic basaltic volcaniclastics are of hydroclastic origin. Deposits either reflect hydromagmatic explosions from direct interaction of water with magma in volcanic vents (*eg.* surtseyan tuff cone at Capel Hill; Chapt 5.4.1.2) or non-explosive fragmentation of subaqueously erupted basalt lava (*eg.* hyaloclastite associated with pillow basalts beneath Lower Rhyolitic Tuff on Gribin Ridge; Fig.3.12, 5.49). In both types, glass and lithic fragments are extensively altered with total replacement by chlorite and/or calcite. Tiny plagioclase laths are the only signs of the original volcanic texture (Fig. 3.12).

3.2.2 SECONDARY PYROCLASTICS

Secondary pyroclastic units include sediment reworked by shallow marine processes (*eg.* top of Garth tuff), as well as material reflecting redistribution by gravity mass movements on unstable depositional slopes (*eg.* 4th volc. mbr. of CCVF). Both types are enriched in crystal and lithic components in relation to their primary counterparts. Although a few vitroclastic (≈ 20% phenocrysts) primary ash flow tuffs occur within the 4th member of the Capel Curig Volcanic Formation (Fig. 3.15), most deposits comprise crystal-rich debris flow deposits (Fig. 3.13) with subhedral to euhedral plagioclase or more rarely quartz crystals and rhyolitic lithic fragments set in a finely crystalline vitroclastic (now a quartz-sericite aggregate) matrix (Fig. 3.16). Juvenile material is often diluted by

incorporation of accessory lithic fragments (aplite) or locally derived siltstone or sericitic chert fragments.

The composition of pyroclastics reworked by shallow marine processes, in contrast, is strongly influenced by the nature of the subjacent deposit (Fig. 3.14). Sorting of sediment by grain size and hydraulic equivalence tends to separate vitric from denser crystal/lithic components (see Roobol, 1976). Shallow marine reworking will abrade fragile cuspate glass shards more readily than crystals (Fig. 3,17) and will eventually lead to physical breakdown and removal of most glass fragments. Quartzose, crystal-enriched sediments deposited 20m above the Garth tuff (section A2, 28m, sample A51; Fig. 3.18), for example, contain scant evidence for the underlying ash flow from which they were derived. The only surviving indication of this volcanic event which produced 25km^3 of ash-grade material is a few scattered angular outlines of glass shards visible in plane-polarised light. In contrast to the above, volcaniclastic sediments immediately overlying the Pitts Head Tuff on Bochlwyd Ridge are dominated by subangular lithic fragments of partially lithified to welded ash flow tuff, large glass fragments with perlitic fracture, and minor rounded quartz crystals set in a chlorite/sericite matrix (Fig. 3.19, 3.20). Shard fabric is only visible in larger clasts; sand-sized grains are optically indistinguishable from chert.

An additional example of textures observed within secondary pyroclastics is provided by the Vuggy Mbr. of the Ddeugwm Formation (Fig. 3.21a). The consistency of composition and grain size over a 10-15m interval, the repetitive occurrence of discontinuous ashfall tuff beds (5-15cm thick), and a direct relationship to subjacent beds of tuffaceous siltstone all provide evidence for penecontemporaneous volcanicity. In thin section, however, the tuffaceous sandstones show no evidence for a volcanic origin; glass shards are completely replaced by a quartz-sericite

aggregate (Fig. 3.21b) with no ghost fabric visible in plane polarised light.

This information on the preservation of volcanic textures within reworked pyroclastics has allowed a genetic interpretation to be given to the white weathering "tuffites" or tuffaceous siltstones (Fig. 3.2b, 3.22) which form an integral part of sequences within the Upper Cwm Eigiau Formation above the Pitts Head Tuff. They have been variably interpreted as intrusive rhyolites (Diggens & Romano, 1968), waterlain tuffs (Wright, 1979, p.282), or more specifically fine airfall into shallow water where they were redistributed by penecontemporaneous erosion (Howells *et al*, 1978, p.16). "Tuffite" horizons are 4-10m thick and organised into 20-50cm thick, often graded, lensoid beds. All are clay-grade rocks. Geochemical analyses (Fig. 2.15, 2.16) indicate at least 78% SiO_2 in most tuffites with some examples containing as much as 83% SiO_2 (F3 tuff). Ash beds were originally assumed (this study) to represent redeposition of the unwelded top portion of the Pitts Head Tuff. The chemical and physical homogeneity of tuff beds, the thickness of the deposits, the complete lack of epiclastic material, and most importantly, the abundance of glass shards (Fig. 3.23 to 3.26) which do not survive marine reworking within secondary pyroclastics (see above), however, all point to a primary pyroclastic origin with NO storage of sediment before deposition. Each "tuffite" horizon is now interpreted to represent entry of vitric-rich pyroclastic flows into the sea. Fuller descriptions and interpretation will be provided later (Chapt. 7.26, 7.4).

The previous discussions highlight both the difficulty in recognising volcanic material in thin section within either primary or secondary pyroclastics, as well as the necessity of detailed petrographic examinations. Problems encountered in assigning an identity to a 4-40 m

thick deposit are small in comparison with those experienced in trying to identify a 0.001 m wide sand grain (Fig. 3.27).

3.3 SANDSTONE PETROGRAPHY: EPICLASTICS or REWORKED PYROCLASTICS?

3.3.1 INTRODUCTION

The detrital composition of a sandstone can be related to several variables (1) provenance (*ie.* composition and relief of source area), (2) climate, (3) modification during transport and deposition and (4) alteration during diagenesis (Suttner, 1974). Medium-grained or coarser (» 0.25mm) sandstones were chosen to minimize possible compositional variations caused by grain size, and maximize petrographic information obtained from lithic fragments. Sandstones were selected if possible from alluvial fan or braidplain settings to reduce the influence of varied sorting parameters between different depositional environments on sandstone petrography. All sandstones point counted are tabulated in Figure 3.28. 300 points (including matrix) were counted from each thin section and were divided into 27 categories (Fig. 3.29). Lithic components (regardless of size) were categorized into the grouping of the larger lithic fragment rather than assigning component sand-sized crystals and grains to the category of the grain itself (Gazzi-Dickinson method: Ingersoll *et al*, 1984). Although the latter approach apparently reduces the influence of grain size on detrital modes, counting genetically significant rock fragments (*eg.* quartz porphyry) as monocrystalline grains (quartz) gives misleading views of sandstone petrography and disguises evidence regarding the true lithology of source rocks (see Suttner & Basu, 1985; Decker & Helmold, 1985; Ingersoll *et al*, 1985). Raw petrographic data and results are presented in Figure 3.30 to 3.32. Petrographic implications of individual crystal and lithic fragments will be discussed first, before examining the regional and stratigraphic variation in the nature of any volcanic input.

3.3.2 DETRITAL COMPONENTS

3.3.2.1 QUARTZ

Quartz is surprisingly common within volcaniclastic sediments (Fig. 3.34), forming on average 54% of the detrital framework (Fig. 3.31, 3.32), ranging from 12 to 93%, and consisting of unstrained (Qm) and strained (Qms) monocrystalline quartz, polycrystalline quartz (Qp), chert (CHERT) and chalcedony (CHALC).

Monocrystalline quartz comprises 24%, of clastic detritus. These grains often display embayed margins, with most embayments being filled by either fine-grained microcrystalline quartz (Qpm) or a quartz-sericite aggregate (QS; Fig. 3.35). One can appeal to several contrasting processes for the origin of this texture. These are as follows: (1) solution pits or embayments can form on quartz sand within tropical soils (Cleary & Connally ,1971), (2) the invasion of quartz by calcite cement during burial diagenesis (*eg.* Shenhav, 1971) and (3) embayment of quartz grains within magma bodies before their eruption (Pettijohn *et al* , 1973). In this case, the absence of any early rims of clay cement on quartz grains (see Suttner and Dutta, 1986), and the short distances of travel (« 10 km) experienced by most detritus (see Chapt. 3.4.1), both suggest that the influence of weathering on quartz shape was minimal. Any calcite cement which formed within volcaniclastic sandstones (Chapt. 3.3.4) was late-stage from the diagenetic alteration of plagioclase, and is not thought to be important in the genesis of the embayments. The above relationships point to a volcanic origin for the embayed quartz, a conclusion supported by the fact that these quartz grains also commonly display a euhedral β- quartz crystal habit (Fig. 3.36, 3.37). The fracturing within subhedral unstrained quartz crystals (Fig.3.38, 3.39) would reflect the 0.3% change in the c/a axial ratio during inversion from β- to α-quartz crystal habits on rapid cooling.

Quartz overgrowths are often present on the monocrystalline grains, particularly within quartz rich sediments. Silica is thought to have been largely supplied either by intrastratal solution of feldspar crystals (Chapt. 3.3.2.2) and/or ferromagnesian minerals, or by solution and reprecipitation of fine-grained microcrystalline quartz matrix and glass shards (Chapt. 3.2), with a smaller contribution from pressure solution between adjoining quartz grains.

Strained monocrystalline quartz comprises on average 10.5% of the detrital framework (Fig. 3.32, 3.33), and probably had a source within plutonic basement rocks rather than extrusive volcanic rocks. The ratio of strained/unstrained monocrystalline quartz, however, is not utilised in later discussions as strain can also be acquired from post-depositional structural deformation (*eg.* samples B30 to B39, section B4, crest of Capel Curig anticline) in rocks with low matrix percentages (*eg.* Fig. 3.34) and grain-to-grain contacts between crystals. Polycrystalline quartz (less than 3 crystal units), makes up 7.2% of the clastic grains.

Chert (Fig. 3.27, 3.40, 3.41) is surprisingly common (9.7%) as detrital grains considering the absence of either bedded cherts or carbonate rocks within underlying strata to act as source materials. Knowledge about the alteration products of primary pyroclastics (Chapt 3.2), however, suggests that most "chert" represents diagenetic alteration of aphanitic rhyolite lithic fragments or vitric-rich volcanic ash. Arguments will later be put forward (Chapt. 3.4.2) supporting the first origin. The genesis of the sericitic chert fragments is even less certain, for they could also be derived from erosion of tuffaceous sedimentary rocks. In the following sections, all chert is included as a rhyolitic lithic fragment (RHY) in any discussions with sericitic chert remaining in the sedimentary lithic category. The uncertainty attached to interpretations of chert genesis, will of course play havoc with any use of

ternary diagrams attempting to discriminate between plate tectonic settings of depositional basins based on sandstone composition.

Lithic fragments of chalcedony (Fig. 3.39, 3.42) are rare as detrital grains. Their occurrence is significant, however, as they are usually found associated with monocrystalline quartz and euhedral sanidine crystals and probably represent devitrified volcanic glass.

3.3.3.2 FELDSPAR

Feldspars comprise approximately 16% of the detrital framework, with the plagioclase to total feldspar ratios (PF) of the 92 samples ranging from .09 to .87 and having a mean of .60. Plagioclase grains are usually subhedral or euhedral (Fig. 3.43, 3.44) and often 2-3mm in length. Reconnaissance EDS analysis suggests most crystals are of albite composition. K-feldspar is dominated by anhedral to subhedral untwinned orthoclase crystals, and euhedral sanidine crystals with well-developed simple twinning (Fig. 3.45).

Most sandstones show evidence for post-depositional albitization of K and Ca components of feldspars. Selective removal of central portions of some plagioclase crystals (Fig. 3.46) is suggestive of crystal zoning (*cf.* Chan, 1985) where the Ca-rich cores were more prone to leaching. Albite with "chessboard" twinning (Fig. 3.47, 3.48) characterized by a blocky to lath-shaped pattern produced by short twin lamellae, probably originated by diagenetic alteration of K-feldspar crystals, with albite produced nucleating on submicroscopic relicts of unreplaced sanidine or orthoclase crystals (see T. Walker, 1984). Most of the Na needed was probably derived from alteration of nearby plagioclase grains; K liberated was usually reprecipitated locally as authigenic K-feldspar cements (*eg.* Fig. 3.88-3.90) whereas Ca was incorporated within authigenic sphene or Mn-rich calcite cement. These "chessboard" albite crystals were originally misinterpreted as crystals of plagioclase during point-counting; visual

estimates were later used to correct feldspar percentages where necessary; the plagioclase/total feldspar ratio is actually lower than calculated (probably ≈ 0.55).

3.3.3.3 LITHIC FRAGMENTS

Lithic fragments including "chert" comprise 39% of the detrital framework and are dominated by volcanic rock fragments (65.1% of the total lithic population). A complete spectrum of rhyolitic hypabyssal and extrusive volcanic rocks can be found (61% of the Lv total) which include, in decreasing order of abundance: (1) Felsitic microcrystalline quartz fragments (chert), (2) Porphyritic rhyolites and/or rhyodacites with equant phenocrysts of quartz, plagioclase and sanidine. (Fig. 3.49, 3.50) (3) Aplite and microgranite with micrographic textures. (Fig. 3.51, 3.52).(4) Glassy rhyolite with aphanitic or microlitic texture (Fig.3.53 - 3.55). Few definitive clasts of rhyolitic tuff (Fig. 3.56) were identified within most epiclastic sandstones examined. Basaltic volcanic rock fragments are poorly preserved, with clinopyroxene or olivine rarely recognized (Fig. 3.57 - 3.61) They are identified by their seriate texture with microlites or more rarely phenocrysts of plagioclase set in a chloritic groundmass. Lithic fragments of chlorite are also inferred to have a basaltic origin (Fig. 3.62, 3.63). Although basalt clasts occur only in small proportions within most depositional systems studied, they comprise the dominant type of volcanic clast in sandstones subjacent to the Pitts Head Tuff at Moel Hebog and Rhyd Ddu (Cwmystradllyn, Y Braich and Cae'r gors mbrs. of Cwm Eigiau Fm.) as well as within sediments of the Magnetite Mbr. of the Tryfan Formation. Other volcanic fragments recognized include trachytic ?rhyolite and ?trachyandesite fragments of intermediate composition (1.7%).

Sedimentary lithic fragments comprise 13% of the detrital framework and are most common in alluvial fan sandstones from the block-faulted NE margin of the basin (*eg.* Tryfan Fm. at Craig yr Ysfa, see Chapt. 4.3.2.4,

4.3.4.5), or from units such as the Racks Mbr. of the Capel Curig Volcanic Formation. They are dominated by clasts of sericitic chert (Fig. 3.64), which probably represent badly altered vitric-rich volcanic ash. Other lithologies include extremely poorly sorted heterolithic clasts of tuffaceous (rhyolitic) sandstones and siltstones (Fig. 3.65 - 3.67); the texture and composition of these sandstone fragments usually mimics that of the sandstone which they are contained within. Plutonic igneous lithic fragments comprise 1% of the detrital framework, and consist of coarsely crystalline granite or granodiorite clasts (Fig. 3.68 - 3.70).

3.3.3 HEAVY MINERALS

Thick (1-2cm wide) concentrations of black-weathering, magnetic, opaque minerals commonly occur either along the tops of foresets of trough cross-stratification (Fig. 3.71) or associated with zones of horizontal upper flow regime lamination (Fig. 3.72). Although characteristic of fluvial or shoreface sequences, the heavy mineral laminae are also found within offshore storm beds (section C3, 96m or 107.3m; Chapt 4.3.4.7; Fig. 4.53). Early emphasis in this research was placed on identification of heavy mineral suites in the hope of aiding provenance studies. It soon became clear, however, that both the lack of detailed petrographic data from subjacent strata and the absence of preserved Ordovician source terrains to the west and northwest, made this impossible. The results provided here, represent data from a few (less than 10) sandstones selected before this line of research was abandoned. Heavy minerals were identified through use of the EDS analyzer in conjunction with the back-scatter SEM[1]. The composition of three fluvial units with particularly notable concentrations of opaque heavy minerals was examined in detail. Relative amounts of heavy minerals were estimated by taking photographs of heavy

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[1]System used was a JEOL JSM-840 SEM and a TRACOR NORTHERN TN-2000 EDS available at Oxford Polytechnic.

mineral bands within these samples at low magnification and identifying all grains contained therein. Disaggregation of sandstone samples to separate heavy mineral suites was not attempted as preliminary investigations indicated that the common occurrence of authigenic overgrowths would mask the true identity of the detrital grains.

In thin section, heavy minerals are found either as discrete horizontal or low angle laminae (Fig. 3.73), or scattered grains "peppered" throughout the sediment (Fig. 3.74, 3.75). The latter mode of occurrence reflects rapid rates of sedimentation with little sorting. Further textural evidence for rapid accumulation is that heavy minerals are commonly NOT in hydraulic equivalence with the surrounding light fraction (Fig. 3.76).

The heavy mineral suite, which comprises on average 3.7% of the detrital framework (among the 92 sandstones point counted), is dominated by opaque minerals. Magnetite is the most common species comprising 58% of the detrital grains (n=770) identified in thin section (Fig. 3.77). Although isolated accumulations of magnetic heavy minerals occur "stranded" within thick sequences of "normal" volcaniclastic sediment (*eg.* sample C27, C71), the most notable field occurrences are associated with sediments rich in basaltic detritus (*eg.* Magnetite Mbr., Cae'r Gors Mbr.). Such magnetite-rich depositional systems are in fact responsible for some observed aeromagnetic anomalies in this portion of Snowdonia, as recently confirmed by ground survey (R. Evans, BGS Keyworth, pers. comm., 1986).

Most detrital magnetite has a raggy "sawtooth" outline, with internal exsolution lamellae occasionally observed (Fig. 3.78, 3.93). Geochemical analyses indicate that magnetite commonly has high titanium (15-20% Ti) or manganese contents (3-5% Mn) (Fig. 3.80); a titanium content of 5% was used to separate magnetite crystals into two populations (Fig. 3.77). Although Ti-rich magnetite is a common phase in rhyolitic volcanic rocks (*eg.* Taupo volcanic zone, Hamill & Ballance, 1985), the high titanium contents in

conjunction with ilmenite again favours derivation from mafic volcanic rocks (Buddington & Lindsley, 1964). The low oxide totals calculated from EDS analysis using a magnetite stoichiometry suggest much magnetite ($FeO.Fe_2O_3$) has altered to maghemite (μ-Fe_2O_3) (A.T. Kearsley, pers. comm., 1986) (Fig. 3.80) , and replacement by sphene is occasionally seen in thin section (Fig. 3.81 - 3.83). In spite of this, magnetite (particularly Ti-poor) appears to be chemically more stable than other iron oxides.

Magnetite also occurs as an authigenic phase. Euhedral overgrowths on raggy grains of ilmenite are particularly common (Fig. 3.84, 3.93). Scattered authigenic crystals without any detrital core are less prevalent. They are the dominant heavy mineral species, however, in sandstones immediately below many parataxitically welded ash flow tuffs (Fig. 3.85, 3.86) and probably reflect later ?hydrothermal migration of pore fluids along the base of the impervious tuff sheets. Magnetite is also common within volcanogenic vein mineralization associated with the Lower Rhyolitic Tuff caldera (Reedman *et al*, 1985) and the two types of occurrence are probably co-genetic.

The origin of fine-grained (mud-sized; most less than 3 microns) bioturbated magnetite crystals (Fig. 3.87-3.90) within intermediate depth mouth-bar (deltaic) sequences is more problematic. The common euhedral shape argues against a detrital origin. There are several contrasting processes which could produce these crystals by authigenic means. Firstly, the extensive alteration and subsequent compaction of ilmenite or titanomagnetite could result in a loosely packed, poorly crystalline matrix (protomatrix) of leucozene and iron oxides (Morad & Aldahan, 1986). In contrast, several authors (Kirscavink & Chang, 1984) have recently interpreted ultrafine-grained magnetite (« 0.5 μm) in deep sea sediments to represent bacterial magnetofossils formed at the mud-water interface within oxygen poor (but not anoxic) sediments. Although the distribution of

magnetite (Fig. 3.88) favours the latter mechanism, it is not certain whether suitable organisms were available to catalyze precipitation of magnetite of the size observed within Ordovician deltaic sediments. It is intriguing, however, that the distribution of marine fauna in North Wales and lower Palaeozoic successions in general (*eg.* Cave, 1965; Berry & Wilde, 1978; Brenchley & Pickerill,1980) also suggests poor ventilation of Ordovician epicontinental seas with a tendency for individual sedimentary basins to stagnate and become depleted in oxygen.

Ilmenite is the other common heavy mineral within sandstones and comprises 12% of the total population (Fig. 3.91 - 3.94). Most grains show extensive diagenetic alteration, with TiO_2 (rutile, leucozene) and sphene the most common products. Iron liberated from the progressive enrichment of titanium is usually reprecipitated locally as authigenic magnetite overgrowths. The ilmenite (and titaniferous magnetite) appears chemically unstable in comparison with titanium poor magnetite and preferentially occurs in depositional systems directly associated with basaltic source rocks (*eg.* Magnetite Mbr., Fig. 3.77, 3.91).

Other heavy minerals include zircon (4%), garnet (7.1%), rutile (2%), sphene (2.6%), clinopyroxene (1%) , haematite (1%), allanite (trace), and chromite (one occurrence). Detrital zircon crystals (Fig. 3.95 - 3.98) are commonly both idiomorphic and chemically zoned , with embayed margins, features all indicative of volcanic source terrains (*cf.* Callendar & Folk, 1958). Phosphates and oxides rich in Th, La, Nd, Yb, Y, and Di which infill fractures associated with zircon and allanite crystals indicate limited mobility (30 microns) of exsolved light and heavy rare earth elements during alteration. The garnet population is dominated by species which suggest a felsic (rhyolite,granite) parentage (*eg.* almandine, spessartine, rare grossular). Allanite (Fig. 3.99, 3.100) and chromite

(Fig. 3.78), although volumetrically insignificant, imply both siliceous and ultramafic source rocks respectively (Deer *et al*, 1966, p. 69, 432).

In summary, the heavy mineral suite was primarily derived from basaltic sources with secondary contributions from rhyolitic or granitic "basement". The absence of other silicate minerals (*eg.* olivine, hornblende) that one would expect from mafic sources, and the low amount of clinopyroxene found, indicates some chemical weathering before deposition. Comparison of heavy mineral assemblages from the 3 depositional systems examined (Fig. 3.77) indicates progressive dilution of detritus from mafic volcanic sources (titaniferous magnetite, ilmenite, clinopyroxene association) by material derived from the uplifted basement sources (more garnet, zircon, rutile, allanite, no ilmenite). Magnetite persists, however, due to its mechanical and chemical stability. Its abundance in the Capel Curig and Moel Siabod depositional systems, however, is probably overestimated by the selective sampling techniques employed.

3.3.4. MATRIX AND AUTHIGENIC COMPONENTS

Full understanding of early post-depositional changes within Ordovician volcaniclastic sandstones is prevented by the later lower greenschist facies metamorphism. However, some textural relationships and mineral assemblages observable within the matrix and cement facilitate our understanding of the provenance of the detrital component. Although all sediments are held together by both "greywacke" matrix, and externally derived authigenic cement, there is usually a dominance of one of these components within each sandstone, and they will be considered individually.

The matrix component is dominant, comprising 18% of the total sandstone and consisting of either chlorite, quartz-sericite aggregates, or microcrystalline quartz. There is usually a direct correspondence between the composition of the detrital framework and the nature of the intervening matrix, which is well displayed by comparing the Lvb:Lvr:LsLp and Qp:Qs:Chl

plots (Figures 3.118, 3.119). For example, chlorite or haematite matrix are most abundant in sandstones rich in basalt clasts (*eg.* Tryfan Fach (5), Moel Hebog(11), and Cae'r gors members(12)). In contrast, sandstones containing abundant rhyolitic debris (*eg.* Reworked Pitts Head Tuff (15,16) or Upper Cwm Eigiau Fm. (UC)) have matrix dominated by microcrystalline quartz (Qp). Sandstones derived from mixed terrains or pre-existing sedimentary rocks (*eg.* Capel Curig(6) or Moel Siabod(7) depositional systems) contain all three matrix components. Secondary authigenic minerals in matrix rich sandstones include minor authigenic K-feldspar, sphene, magnetite and calcite which are usually intimately associated with the plagioclase, K-feldspar, or heavy mineral from which they were derived. The above relationships a diagenetic origin for most matrix within Ordovician volcaniclastic sandstones (see Brenchley, 1964). This means that the original detrital composition of most grains can be determined by consideration of the surrounding secondary component.

Sandstones which contain abundant quartz or calcite cement are usually poor in the above matrix components and rich in quartz (Fig. 3.101, 3.102). Authigenic quartz preferentially forming on unstrained monocrystalline quartz crystals. Both cements are probably of early origin; by reducing the permeability of sandstones they prevented matrix-producing reactions from proceeding during later burial diagenesis.

3.4 IMPLICATIONS OF PETROGRAPHY

3.4.1 CONTROLS ON SANDSTONE COMPOSITION

The volcaniclastic sandstones are dominantly lithic arenites (Fig. 3.103) with an average composition of $Q_{44}F_{19}R_{39}$ (with chert as Lv). The common occurrence of embayed monocrystalline euhedral β-quartz, euhedral plagioclase, sanidine, or orthoclase, and rhyolite or basalt lithics all attest to a volcanic provenance. The variation in sandstone petrography is striking; the amount of quartz, for example, ranges from 12-78 percent. As

a result, full appreciation of the volcanic contribution to the basin, can only be achieved by consideration of each depositional system individually. Sandstones were grouped into 17 petrographic units corresponding to individual depositional systems or volcanic events (Fig. 3.104a,b). For comparative analysis of the provenance of sandstone suites, detrital framework grains were recalculated into the volumetric proportions of the categories of grains indicated in Figure 3.105a; the mean composition of Caradocian volcaniclastic sandstones recast into these grain categories is tabulated in Figure 3.105b. A series of complementary triangular diagrams, each involving a different set of grain populations was used to display data. First, however, it is fruitful to comment on other possible controls on sandstone composition. These include climate and weathering, depositional environment, grain size, and burial diagenesis.

It is most difficult to assess the full role played by climate. Evidence from depositional sequences (*eg.* rarity of desiccation cracks, unexpected absence of debris flows which are usually characteristic of volcanically-influenced settings), combined with palaeomagnetic evidence (*eg.* Piper, 1978), suggest a humid (*ie.* abundant water supply), sub-tropical climate. Basalt, volcanic glass, and feldspar crystals will be particularly susceptible to chemical weathering; more attention must be paid to the character and amount of interstitial matrix and cement within sediments thought to have been derived from source rocks rich in these components. It is possible that the quartz-content of some sediments (particularly those derived from pre-existing sediment) deposited during tectonically quiet periods, with slower rates of deposition (such as the Gwern Gof (2), Moel Siabod (7), or Racks Members (9)) could be enhanced through selective removal of more unstable lithic or feldspar fragments (*cf.* Suttner & Dutta, 1986; Franzinelli & Potter, 1983). Most quartz-rich sandstones in this study, however, were deposited during tectonically

active periods (*eg.* Bochlwyd Mbr. (13), Upper Cwm Eigiau Formation (UC)) and transported extremely short alluvial distances (less than 10 km) in high energy braided streams from "youthful" fault blocks (Chapt. 4). The extremely high rates of sedimentation indicated suggest little storage of sand on bars or floodplains to permit multiple episodes of weathering within the same sedimentary cycle. Although heavy mineral suites are dominated by stable minerals such as magnetite and zircon they do contain notable quantities of more unstable minerals such as pyroxene and garnet. The above factors, combined with the extreme variation in composition displayed by sandstones, mitigate against a strong influence of climate and surface weathering processes on detrital mineralogy.

The influence of depositional environment and transport history on sandstone composition can be seen by examination of the lateral variation in petrography of sediment within alluvial fan-braidplain-deltaic depositional systems subjacent to the Capel Curig Volcanic Formation (petrographic units 4-7), as shown both on rose diagrams of individual variables (Fig. 3.105) or QFR and QmFLt ternary plots (Fig. 3.106 - 3.109). Alluvial depositional systems, as expected, generally display a downstream increase in the amount of monocrystalline quartz, increases in feldspar content, and decrease in relative amount of lithic fragments; the magnitude of these changes for the Capel Curig and Moel Siabod depositional systems from Craig yr Ysfa to the Capel Curig Anticline, a distance of 10km (as the stream flowed), are respectively $Q_{38}F_{12}R_{50}$ to $Q_{54}F_{22}R_{25}$, and $Q_{35}F_{7}R_{57}$ to $Q_{60}F_{17}R_{23}$. This variation largely stems from differences in the relative mechanical and chemical stabilities of the various crystal and lithic components (see Kuenzi *et al* 1979 for modern example). The loss of sedimentary fragments upon transport from alluvial fan to braidplain environments is particularly noted, with sandstones displaying a 10 to 40% increase in the proportion of volcanic to sedimentary lithics; the

rhyolitic rock fragments appear more durable and often show no change in absolute abundance over these transport distances. The lateral variations, however, are not as simple as suggested above, and several petrographic changes (see below) necessitate the postulation of additional volcanic sources (Fig. 3.128) feeding the axial NW-SE braidplain. Comparison of sandstone suites from neighboring alluvial fans (parallel to depositional strike) along the NE margin of the basin at Craig yr Ysfa and at Llyn Dulyn, separated by a distance of 3 km, also indicates significant variation in petrography. The higher quartz contents of the latter would reflect both a more pronounced influence of rhyolitic intrusions and a multicyclic transport history of sediment (the sequence at Llyn Dulyn is condensed compared with Craig yr Ysfa: see Chapt. 6.2.2).

The extreme influence that grain size can have on the composition of sandstones within a single depositional system is well illustrated by fluvial sediments from the Magnetite Mbr. of the Tryfan Formation (Fig. 3.10), with samples collected over a total stratigraphic thickness of only 13m. Very coarse sandstones consist largely of aphanitic and slightly porphyritic rhyolitic rock fragments, euhedral sanidine crystals and subhedral plagioclase crystals set in a matrix of microcrystalline quartz (Fig. 3.43). Very fine to coarse sandstones (the dominant lithologies), on the other hand, are dominated by basaltic lithic fragments, large euhedral plagioclase crystals, and abundant opaque (magnetite, ilmenite) or transparent (orthopyroxene, clinopyroxene) heavy minerals set in a chloritic matrix (Fig. 3.75, 3.76). The different compositions reflect the contrasting mechanical and chemical stabilities exhibited by rhyolite and basalt lithic fragments, with chemical weathering in particular quickly reducing the latter to sand-sized detritus.

Finally, the relatively rapid diagenesis of volcaniclastic sandstones, with early dissolution of unstable glass, crystal and lithic components,

could also invalidate interpretations of provenance based on detrital mineralogy (Walker *et al*, 1978). Within Plio-Pleistocene non-marine volcaniclastic sandstones of the Cayagan Basin, Philippines, dissolution of up to one-half of framework grains (plagioclase, heavy minerals, andesite lithics) through hydration reactions has increased porosity to as much as 40% (Mathisen, 1986). Sandstones were altered in a hydrologically open system with removal of dissolved material by pore fluid circulation. Mathisen (*op. cit.*) postulated that such processes could enrich the quartz content of volcaniclastic sandstones.

In North Wales, diagenesis resulted in albitization of k-feldspar, and dissolution and replacement of plagioclase, basaltic lithics and heavy mineral grains. The high heat flow which characterizes marginal basins would assist the hydration reactions. Matrix, however, comprises only 20% of most sandstones. The close relationship between the composition of detrital clasts and distribution of authigenic minerals suggests restricted flow of pore fluid through sediment, rapid occlusion of original permeability and closed hydrological systems. The limitations thus imposed on the degree of diagenesis, with the original detrital composition of most grains (except glass) still recognizable, means that sandstone composition can be used for provenance and plate tectonic interpretations if the influence of the other controls can be moderated.

In this study the influence of transport history and depositional environment was decreased through selection of samples from all portions of each depositional system with restriction to the alluvial component if possible. The role of grain size was reduced by including samples of available size fractions (from medium sand to granule conglomerate) in data sets for each depositional system. If one accepts that climate was an unchanging variable throughout the time span encountered, that chemical weathering did not have time to significantly modify the original detrital

mineralogy, and that the detrital signature was not modified beyond recognition by later diagenesis, then the petrography of sandstones can be used with caution to monitor stratigraphic and regional changes in the nature of source rocks within the basin.

3.4.2. THE VOLCANIC CONTRIBUTION

Results of the petrographic analyses, which aim to document the volcanic input in particular, are tabulated in Figure 3.111, and displayed in a series of ternary diagrams (Fig. 3.112 to 3.119) and plots of the plagioclase/total feldspar and lithic volcanics/total lithics ratios (Fig. 3.120). Brief discussions of selected sandstone suites are presented below with numbers in brackets referring to the petrographic unit(s) described.

The Gwern Gof Mbr. of the Ddeugwm Formation (2), the Moel Siabod Mbr. of the Tryfan Fm. (7), and the Racks Mbr. of the CCVF(9) consist of poorly to moderately sorted medium sandstones (Fig. 3.111- 3.113). The fact that most grains are anhedral, the high percentages (22%; 24%; 35%) of matrix, and the low contents of obvious volcanic lithics (Lv/L ratios = 0.24, 0.36, 0.11) with sericitic chert and siltstone comprising most rock fragments (Fig. 3.121), suggest derivation from a source dominated by tuffaceous sedimentary rocks, with the moderate-high quartz contents reflecting several cycles of erosion and/or a long distance of transport. The style of volcaniclastic sedimentation in all three sandstone suites (Chapt. 4-6) supports these findings and indicate deposition during "quiet" tectonic and volcanic intervals. The slightly greater proportion of rhyolitic rock fragments in the Moel Siabod Mbr., as well as the common occurrence of outsized quartz crystals, reflects the diminishing influence of the quartz porphyry terrain which was active during sedimentation of the Capel Curig Mbr. (see below).

4) The genesis of sandstones within the Tryfan Fach Mbr. is enigmatic. Alluvial fan sediments at Craig yr Ysfa ($Qm_{13}F_{24}Lt_{63}$:

$Qp_{17}Lv_{43}Ls_{40}$) contain significant monocrstalline quartz (27.2%), little feldspar, and lithic fragments dominated by aphanitic rhyolite or sandstone (13%), with 3% of granitic or rarely gneissic lithic fragments (Fig. 3.104b). Although sandstones are poorly sorted, most grains are rounded. The composition and texture indicate derivation from a mixed plutonic basement/tuffaceous sediment/rhyolite volcanics source. In contrast, sandstones within correlative braidplain sequences to the south ($Qm_{13}F_{24}Lt_{63}$: $Qp_{16}Lv_{93}Ls_{1}$) contain large plagioclase and/or sanidine crystals, a lower percentage of monocrystalline quartz, and lithic fragments dominated by porphyritic rhyolite fragments with minor clasts of basaltic lava(Fig. 3.122). Intercalated airfall tuffs indicate contemporary silicic volcanism. The amounts of both basaltic detritus and primary rhyolitic ashfall increase to the southwest (compare samples C56, C8 on Fig. 3.31). The composition of braidplain sandstones suggests dilution of the Craig yr Ysfa source by crystal-rich rhyolitic and minor basaltic volcanics and indicate penecontemporaneous pyroclastic volcanism of largely rhyolite composition to the west or southwest.

5) The Magnetite Mbr ($Qm_{13}F_{24}Lt_{63}$: $Lvb_{27}Lvr_{67}LsLp_{7}$) displays considerable variation in composition according to the grain size of the sediment, as discussed previously. The sudden occurrence of relatively undiluted basaltic detritus into the basin indicates uplift and erosion of a sequence of basaltic lava and tuff; penecontemporaneous volcanicity is suggested by the uniformity of sandstone composition within very fine to coarse sandstones.

6) The Capel Curig Mbr. depositional system displays considerable regional variation in composition (Fig. 3.106 to 3.109). Alluvial fan sediments at Craig yr Ysfa ($Qm_{13}F_{24}Lt_{63}$: $Qp_{17}Lv_{43}Ls_{40}$) , with the lithic component dominated by fragments of siltstone and sandstone, suggest derivation from tuffaceous sediment. This contrasts with the quartz-rich

aspect of volcaniclastic sandstones from the Capel Curig anticline ($Qm_{13}F_{24}Lt_{63}$: $Qp_{17}Lv_{43}Ls_{40}$; petrographic unit 6D, samples B30-B39) in which the rock fragments consist largely of rhyolite or aplites. Depositional facies (Chapt 4.3.4) suggest sediments were deposited in high energy nearshore environments on a wave-influenced braidplain delta. Although embayed quartz grains are not destroyed within at least moderate (100km) distances of transport within low energy fluvial systems (Franzinelli & Potter, 1983), it is difficult to envisage similar preservation within high energy braided streams and nearshore environments with significant wave power, particularly where the quartz grains are also fractured from the β to α quartz transformation. The poor sorting of sandstones, and nature of depositional facies, however, suggest a limited transport history of sediment within the shallow marine environments and the sandstone petrography is interpreted to directly reflect the characteristics of the source area. This implies introduction of quartz-rich detritus onto the braidplain from an unpreserved porphyritic rhyolite or rhyolitic ash terrain. Facies analysis of depositional sequences (Chapt 4.3.4.8) suggests that alluvial fans supplying such material probably drained uplifted fault blocks to the SW of Llanberis pass (Fig. 3.128).

The composition of the Moel Hebog and Cae'r gors members of the Lower Cwm Eigiau Fm. in SW Snowdonia (Fig. 3.111; 3.128; enclosure 1; $Qm_{23}F_{17}Lt_{59}$: $Lvb_{39}Lvr_{45}LsLp_{16}$ & $Qm_{34}F_{22}Lt_{16}$: $Lvb_{26}Lvr_{56}LsLp_{18}$) contrasts strongly with the petrography of sediments deposited in the northeast portion of the study area. They are characterised by low percentages of quartz, with lithic fragments including significant basalt and mudstone in addition to the ubiquitous rhyolite component (Fig. 3.124). Abundant opaque minerals derived from basaltic detritus (Fig. 3.125) were reworked on wave dominated fan deltas into a series of shore-parallel nearshore bars. Analysis of depositional sequences indicates derivation of sediments

from the west and northwest; the NNE-SSW Cwm Pennant fracture is implicated as the source of the basalt magma (Fig. 3.128). Mudstone fragments are probably lower Caradocian in age (*ie.* equivalent to Nant Ffrancon Fm.).

The fluvial sediments of the Bochlwyd Mbr. (Fig. 3.126, 3.127a), in contrast, consist largely of either anhedral monocrystalline quartz (55%) or rhyolite rock fragments (23%) set in a calcite/quartz cement. Palaeocurrents (*eg.* Fig. 4.78) indicate derivation from the south; the quartz-rich composition implies a short lived renaissance of the porphyritic rhyolite source (at Cwm Dyli, see below) which had earlier supplied material to the Capel Curig braidplain.

The Upper Cwm Eigiau Formation (petrographic unit 17; UC on diagrams) represents the culmination of Caradocian volcaniclastic sedimentation in North Wales. As previously noted, it contains significant thicknesses of tuffaceous siltstones now interpreted to represent primary ash flow tuffs. The intercalated epiclastic component (*eg.* Fig. 3.36) containing abundant monocrystalline quartz (43%), a lithic population consisting entirely of rhyolite (10% porphyritic, 10% aphanitic "chert") or rhyolitic pyroclastics (sericitic chert) and a matrix of microcrystalline quartz is thought to be derived entirely from rhyolitic volcanic sources.

The above analysis of the provenance of various stratigraphic units has resulted in several unexpected revelations regarding the extent, variability, and nature of the volcanic control on sandstone petrography. The influence of a granitic basement, uplifted fault blocks of tuffaceous sediment, and rhyolite volcanic/volcaniclastics was perhaps expected considering the known regional context.

The paucity of obvious clasts of welded or even unwelded ash flow tuff among this detritus, however, is surprising. This implies that major pyroclastic eruptions of rhyolitic composition were largely preserved in situ (*eg.* Garth tuff, Pitts Head Tuff). The large fluidal ash flow tuffs

would preferentially follow lowland areas and thus reduce the likelihood of later erosion during fault block movement. The tectonic changes associated with such volcanicity (Chapt. 6) would ensure preservation. High-level rhyolite intrusions, in contrast, by the very nature of their origin and emplacement would more often than not outcrop in uplifted areas and provide the rhyolitic lithic fragments. Small, ?more viscous, and unwelded pyroclastic eruptions would also be prone to erosion and probably provided the ubiquitous sericitic chert component.

The discovery of a hidden source of basaltic detritus within the basin is perhaps even more important (Fig. 3.128); it has profound implications for the structural evolution of the basin, importance of various fault trends in generating magma, and the tectonic stress regime in North Wales during the Caradoc. Authors (*eg.* Kokelaar *et al*, 1984) had previously speculated that in absence of any signs of major mafic residue, the large volumes of acidic rocks must be derived by partial fusion of underlying crust. This documentation of basaltic magmatism to the west allows an origin of rhyolites by fractional crystallization, in accordance with more recent petrogenetic models (*eg.* Campbell *et al*, 1987; Ball & Merriman, in press).

To grossly simplify the regional variation in the aspect of source rocks, the study area can be divided petrographically into two sub-basins about Llanberis pass, with input of basaltic and mudstone detritus into the SW portion, and a mixed source of tuffaceous sediment, rhyolitic volcanics and a granitic basement feeding the basin in the northeast (Fig. 3.128). For a large part of the evolution of the basin, these two depocentres were segregated by the central uplift of a small rhyolitic quartz porphyry horst block (and small associated centre of siliceous pyroclastic activity) in the vicinity of Cwm Dylli (SH 6454). Besides influencing the petrography

of local sediments, this topographic high guided the entry of the Pitts Head Tuff into the sea (Chapt. 7.3.2.1).

Stratigraphic variation in sandstone composition is just as profound. To generalize the petrographic evolution of the NE sub-basin, there is a gradual reduction in the importance of uplifted granitic/sedimentary fault blocks as sources of sediment. The gradual "patchwork" blanketing of the Caradocian "countryside" by rhyolitic ash flow tuffs and associated intrusions, left this material as the only possible provenance. The Upper Cwm Eigiau Fm. represents the culmination of this trend and records the most prolonged input of volcaniclastic material (both epiclastic and pyroclastic) into the basin. In spite of this, these sediments are among the most quartz-rich of all deposits examined. This is directly at odds with conventional wisdom concerning the effect volcanism will have on sandstone petrography; most authors assume that "highly tuffaceous sands have abnormally low quartz contents [of] less than ten percent in some cases" (Pettijohn *et al*, 1973, p. 235).

3.4.3 PROVENANCE AND PLATE TECTONICS

Although it has long been recognized that the detrital composition of clastic rocks is related to the nature of the source rocks, it is only recently that researchers have attempted to elucidate the relative importance of this control and factors such as transport history, depositional environment, and climate on sandstone petrogenesis.

A semi-quantitative investigation (Garzanti, 1986) of Triassic fluviodeltaic and lagoonal volcaniclastic sediments in the southern Alps found that the effect of different sedimentary processes and hydraulic sorting parameters was never strong enough to obscure the fundamental relationship of sandstone composition to provenance. Although volcanic rock fragments were reduced in proportion by selective destruction within marine environments, the Q/F ratio of the sediment was <u>not</u> affected. It has also

been shown (Suttner & Dutta, 1986; Dutta & Suttner, 1986) that although systematic variations in compositional maturity of first cycle fluvial sandstones in the Gondwana Supergroup (Permian-Triassic intracontinental rift) of India (*eg.* QFR percentages change from 54:42:4 to 99:1:0) are largely a function of changing palaeoclimate, the tectonic imprint on framework minerals remained "indelible" with sandstones continuing to plot within the continental block provenance field of Dickinson and Suczek (1979) (Fig. 3.129). This apparent support for the validity of plate tectonic-sandstone composition ternary diagrams is somewhat misleading, in that if similar increases in the relative amount of quartz/feldspar, with associated decreases in lithic content, were imprinted onto first-cycle "magmatic-arc" volcaniclastic sandstone from an intracontinental rift (*eg.* A, B in Fig. 3.129), the derivative relict volcaniclastic sandstone might plot in a totally different (*eg.* continental block or recycled orogen) tectonic field (see Mack, 1984). As discussed in previous sections, however, the rates of sedimentation, depositional framework, and sandstone petrography all suggest climate was of secondary importance in determining the detrital composition of sandstones.

The relationship between provenance and plate-tectonic setting is more tenuous. The objective is to "use detrital framework modes of sandstone suites [to] provide information about the tectonic setting of basins of deposition and associated provenances" (Dickinson et al, 1983). Compilation of numerous published data allowed data reduction to several petrographic categories of sandstones, distinguishable from one another by location on various ternary diagrams (Fig. 3.127, inset triangles of Fig. 3.112 to 3.116 as well), and corresponding to specific tectonic provenances (Dickinson & Suczek, 1979; Dickinson & Valloni, 1980; Maynard *et al*, 1982; Dickinson *et al*, 1983; Valloni & Mezzadri, 1984).

Several authors have recognized examples of sandstones whose tectonic setting does not coincide with their location on triangular provenance diagrams. Velbel (1985) noted that in accretionary prisms mineralogically mature sands rich in quartz had been transported long distances by turbidity currents from tectonically unrelated source areas. In a situation with more similarity to the Welsh marginal basin, Mack (1984) described the petrography of Tertiary sandstones deposited within the Rio Grande rift during the transition from back-arc basin to continental rift tectonic regimes; sediments deposited within the latter setting were largely derived from relict source rocks of the earlier episodes of arc-related volcanism, with the only evidence for uplifted basement residing in a minor (less than 2%) percentage of granitic rock fragments.

In North Wales, there is no volcanological (see Kokelaar *et al*, 1984) or sedimentological (this study: Chapters 4-7) evidence to suggest any change in tectonic regime **within** the Caradocian. The short distances of transport experienced by all coarser lithologies necessitates very similar tectonic settings for both source areas and depositional basins. Inferences drawn from the provenance of source rocks should relate directly to the tectonic setting of the depositional basin. Although the ternary plots of this study are not directly comparable with those of Dickinson, Ingersoll, and Suczek as a result of the different point counting techniques used, the discrepancy between the two methods is insignificant. If all crystalline rock fragments (granite: 1% of detrital framework at 70% quartz, 30% feldspar) and all porphyritic rock fragments (dominantly rhyolite: 8% of detrital framework with 30% phenocrysts comprising 25% quartz, 55% plagioclase and 20% k-feldspar) were recast into their constituent components, this would only result in a 3-4% decrease in total volumetric percentage of lithics, a 2.5% increase in feldspar contents, and a 1.5% increase in quartz content.

Even accounting for this minor difference, most results from Caradocian volcaniclastic sandstones in North Wales still plot in the "wrong" tectonic field(s) on the widely used QFR or QmFLt diagrams. The mean composition of the 92 sandstones examined (Fig. 3.130, 131), lies at the chert-rich end of the recycled orogen field on the QFR diagram of Dickinson and Suczek (1979), or in between back-arc basins and trailing edge fields on the same plot from Valloni and Maynard (1981). On the QmFLt diagram, the Welsh data are stranded between the recycled orogen and magmatic arc provenances. The dismal failure of plots using total grain populations to recognize any magmatic component stems from the high quartz content, and relative absence of feldspar, resulting from that volcanicity. The expected continental block component (ie. plotting on the quartz-feldspar join) is hidden by the volume of volcanic and sedimentary detritus which dominated the "youthful" fault blocks being uplifted. Individual deposition systems do plot on the ?expected transitional arc field if influenced by basaltic or quartz-poor sources (eg. Moel Hebog Mbr.(11), Magnetite Mbr.(5), 4th member of CCVF (10)); where rhyolitic volcanics (volcanic ash or high-level plutons) are involved as sources, the detrital compositions can plot in either quartzose recycled (Bochlwyd Mbr. (13)) or uninformative MIXED (Upper Cwm Eigiau Formation, (UC)) tectonic fields. Perhaps realizing the compositional variability that one would expect in tectonic settings such as ensialic marginal basins, in their latest paper advocating the use of QFL and QmFLt diagrams, Dickinson *et al* (1983), purposely exclude sandstone suites where the standard deviation on any triangular pole exceeds 10%. Although this method "cleans-up" the data set and results in better clustering of points for each provenance field, it deliberately avoids consideration of marginal basins like North Wales ("standard deviations" on Q, F, and R poles are 16, 10, and 17 percent respectively) which would add "fuzziness" to a oversimplified tectonic

picture. The marginal basins which are included are dominated by monolithological oceanic examples from the Western Pacific, where the majority of sedimentary detritus is derived from adjacent dominantly andesitic volcanic arcs.

Ternary plots using partial grain populations (Ingersoll & Suczek, 1979), emphasizing polycrystalline or lithic components (*eg.* QpLvLs, LpLvLs), and avoiding the "quartz problem" obtain more satisfactory results and appear as better indicators of tectonic setting. Most North Wales sandstone suites fall within the "MIXED MAGMATIC ARC and RIFTED CONTINENTAL MARGINS (Back-arc basins)" fields (Fig. 3.115, 116). The range in sediment composition along the LvLs join, however, is enormous and several petrographic units (1: Vuggy Mbr., volcanic ash altered to sericitic chert and 9: Racks Mbr., derived from tuffaceous sandstones during a tectonic lull) plot in the overlapping suture belt or rifted continental margin settings.

Plagioclase to total feldspar (mean 0.60, range 0.43 to 0.78) and volcanic lithics to total lithics (mean 0.62, range 0.10 to 0.95) ratios of petrographic units are similarly non-committal. Using the values derived by Valloni & Maynard (1981) from the petrography of deep sea sands (Fig. 3.120), the Welsh marginal basin would be assigned to either strike-slip or back-arc basin settings with the mean composition being most similar to the former. Individual samples, however, plot in all tectonic fields.

The amount of research completed to date is insufficient to properly assess whether the breadths of sandstone composition and abundance of quartzose volcaniclastics which characterise the Welsh marginal basin, are found in similar tectonic settings elsewhere. Available petrographic evidence, however, suggest that the observations from this study are not unique, with three notable cases being:

1) the Cretaceous marginal basin in Chile and extending to South Georgia, which developed between the South American continent and an active calc-alkaline volcanic arc (Winn, 1978; Winn & Dott, 1979). Quartz-rich sediment gravity flows (*eg.* Sandebugten Fm. $Q_{35}F_{21}L_{43}$) were derived from siliceous volcanic (quartz and albite phenocrysts in felsite) and sedimentary (quartzose sedimentary fragments) rocks of the Jurassic Tobifera Fm. to the east.

2) Sandstones from the Miocene Tesuque Fm. in the Espanola basin of the Rio Grande rift have been separated into two sedimentary provinces based on the character of lithic fragments (Cavazza, 1986). Those consisting of volcaniclastic and detritus (his province B) derived from the Taos plateau-Latir volcanic field have a higher quartz content (QFL % = 58:31:11) than material eroded from plutonic and metamorphic Precambrian basement (QFL = 51:46:3).

3) A related study (Ekas *et al, 1986)* describing sediments from the southern portion of the same basin, provides an excellent example of errors which can be incurred through routine use of existing ternary diagrams. Sandstones are again divided into two petrographic groups using both field aspect and quartz content. In obvious volcaniclastic sandstones ($Q_{10-31}F_{9-39}L_{30-81}$), the lithic component indicates derivation from an andesitic or dacitic volcanic field. They assume a granitic source for "arkosic" detritus ($Q_{47-66}F_{23-37}L_{9-24}$) due to the greater proportion of K-feldspar and higher quartz contents, even though all lithic fragments and cobbles within this sandstone suite consist of "a combination of microlitic, lathwork, and felsitic" volcanic lithic types", **and** palaeocurrent data suggest derivation from "the compositionally diverse intermediate to silicic volcanic rocks" of the Taos Plateau. In light of the evidence presented, the author questions the validity of this interpretation.

In conclusion, it appears that quartzose volcaniclastic sandstones are a common component of the infill of ensialic marginal basins or continental rifts. The results warn against too facile an interpretation of petrographic data, without considering the regional context. Petrographers either need to broaden the magmatic arc provenance field on existing QFL and QmFLt ternary diagrams and incorporate data from basins such as North Wales, or treat back-arc basins as an anomaly which can be conveniently excluded from such plots because of the heterogeneous nature of their sandstone compositions. If the former approach is adopted, the various tectonic fields will eventually overlap to such an extent, that the original aims become unattainable. The latter approach (adopted by Dickinson *et al* , 1983) is hardly scientific.

At first sight, the plots of Ingersoll and Suczek (1979) using lithic components (QpLvLs; LmLvLs) are more successful. It must be remembered, however, that data were largely collected from deep sea sands. Only time will tell if inclusion of more data, particularly from continental volcanic fields, subaerial and shallow marine sediments, and tropical climates, will also gradually invalidate the apparent discriminatory power of these plots.

The above cautionary notes, however, do not mean that investigations of sandstone petrography itself is worthless. In fact, the results from this study show that full understanding of the genesis, deposition, and tectonic implications of volcaniclastic sequences is only possible through use of sandstone petrography **in conjunction with** facies analysis. Any attempt to interpret a volcaniclastic succession, through the single-minded application of the latter method, will result in as many errors as from sole use of sandstone petrography. Volcaniclastics necessitate the integration of the two contrasting approaches to properly interpret the interplay between tectonics and sedimentation.

Chapter 4 FAN-DELTAS and BRAIDPLAIN-DELTAS: A CONTINUUM OF VARIABILITY

" What's in a name? That which we call a rose by any other name would smell as sweet "

Shakespeare, Romeo and Juliet, II, ii, 43-44.

" In the space of one hundred and seventy six years the lower Mississippi has shortened itself two hundred and forty two miles. That is an average of a trifle over one mile and a third per year. Therefore, any calm person, who is not blind or idiotic can see in the Old Oolitic Silurian period, just a million years ago next November, the lower Mississippi was upwards of one million three hundred thousand miles long and stuck out over the Gulf of Mexico like a fishing rod. And by the same token, any person can see that seven hundred and forty two years from now, the lower Mississippi will be only a mile and three-quarters long "

Mark Twain, Life on the Mississippi, 1884.

4.1 INTRODUCTION TO SEDIMENTOLOGY

Although Ordovician sedimentation in Wales was dominated by graptolitic mudstone, the Caradocian Series contains significant thicknesses of sandstone and conglomerate, frequently containing shelly macrofauna intercalated with the widespread volcanic sequences. In North Wales, several hundred metres of tuffaceous sediment, deposited in a variety of environments, are directly associated with the Llewelyn and Snowdon Volcanic Groups. The area was predominantly the site of marine sedimentation, partly as a consequence of the mid-Ordovician worldwide eustatic sea level rise, which resulted in production of widespread shallow epeiric seas (Fortey, 1984). Most authors still believe "that volcanic islands were **uncommon** and **ephemeral** and that even quite major eruptions commonly caused little or **no change** in the facies and composition of the 'normal' sediments occurring immediately beneath and above the volcanic horizon" (Kokelaar *et al*, 1984, p.265). It had been noted as early as 1838 by Sedgwick, however, that "through wide tracts of country [the volcaniclastic sequences] are unfossiliferous". The sudden occurrence

within the Caradocian of coarser lithofacies, and unquestionable shallow marine fauna, associated with pyroclastic units historically thought to be diagnostic of subaerial volcanicity, leads one to question the validity of the above assumption of an unchanging marine environment.

This study represents the first detailed attempt to unravel the complex depositional history of any of the volcaniclastic successions. Over 10 km of vertical section of epiclastic sediment were examined during field work, with 7200 metres of detailed logs extracted from the coarser lithologies (primarily fine sandstone to conglomerate). Siliceous siltstones also yield valuable sedimentary information as their composition prevented formation of slaty cleavage during structural deformation. The location, vertical sequence, description and detailed interpretations of all sedimentary logs are included in Appendix 5. It is obviously not possible to discuss the reasons behind the interpretations and depositional environments suggested for each metre of sediment. The approach taken, therefore, is to select representative volcaniclastic sequences from both subaerial and subaqueous depositional environments, and interpret these successions in detail. Sequences are later placed in their regional and stratigraphic context, and together with information from sandstone petrography (Chapt.3) used to outline the evolution of the basin related to tectonic and volcanic episodes of varying magnitude and chemical composition (Chapt. 6).

In both "sedimentary" chapters, description of depositional sequences is kept to a minimum within the main body of text, through constant reference to both vertical profiles (either as figures or in the appendix) and photographs. Explanation of symbols and facies codes used in these vertical profiles is shown in Figure 4.1. Emphasis is placed throughout on analysis of facies associations to identify depositional environment rather than detailed inquiries into the origin of individual lithofacies.

Interpretation of coarse-grained fan- and braidplain-deltas, with associated coastline environments, has provided the most useful information regarding possible controls of volcanicity on basin tectonics and related sedimentation. An analysis of most aspects of these depositional systems is presented below, with a shorter examination of the volumetrically more important shallow marine and offshore lithofacies (Chapt 5).

4.2 APPROACH TO DELTAIC SEDIMENTATION

4.2.1 RECOGNITION OF DELTAIC SEQUENCES

Deltas can be defined as coastal accumulations, partly subaerial, built by a river into or against a permanent body of water (Barrell,1912; Wright, 1985). They represent an interaction between sediment laden alluvial systems and basinal processes. In coarse-grained deltas, sediment transport is usually in short, high gradient streams that are often braided, from small drainage basins. The grain size of sediment delivered to the delta front varies from medium sand to large boulders. Resultant sedimentary deposits comprise subaerial, transitional and subaqueous components.

A spectrum of coarse-grained deltaic sequences dominate the Caradocian infill of the marginal basin in North Wales. Determination of ancient delta conditions, which require the demonstration of evidence that "both subaerial and subaqueous sequences were deposited" (Barrell, 1912, p. 383) is not easy where deltaic bodies often prograded into unfossiliferous lacustrine environments. Conglomerates or sandstones interpreted as of fluvial origin, which occur in the upper parts of coarsening-upwards sequences, can be distinguished, however from deltaic or shallow marine deposits by the following criteria:

(1) Typically very poorly sorted with angular pebbles of rhyolite, rhyolitic tuff, siltstone or sandstone in a coarse sandstone matrix.

(2) Commonly rich in heavy minerals with magnetite, ilmenite and

zircon found along upper flow regime horizontal lamination or tops of foresets of cross-stratification.

(3) Lithofacies commonly organised into 2-8m cycles displaying upward decreases in both sediment calibre and the set size of cross-stratification.

(4) Cross-stratification or clast imbrication yield unidirectional palaeocurrents.

(5) Absence of sedimentary structures (*eg.* wave ripples, hummocky cross-stratification) indicative of subaqueous environments.

4.2.2 THE ALLUVIAL COMPONENT

Reconstruction of channel morphology has historically been one of the prime aims of fluvial sedimentology. Modern channels were initially classified either on the basis of channel pattern into (1) straight (2) braided, (3) meandering (after Leopold & Wolman, 1957) and later (4) anastomosing (Rust, 1978) types OR subdivided on their sediment load (*eg.* Schumm, 1981) into: (1) suspended load and dissolved load channels (bedload « 3%) (2) mixed load channels (bedload 3 - 11%) and (3) bedload (bedload » 11%). Experimental work recognised a gross correspondance between type of sediment load moved through the channel and channel pattern (Fig. 4.2a). It has also indicated that pattern changes between braided, meandering and straight appeared to occur abruptly at river pattern thresholds which depended on critical values of stream power, gradient and sediment load (Schumm, 1981). Increases in either of these variables would favour development of braided channel patterns. Debris flows and sheetfloods, which would behave in transport as non-Newtonian fluids, were recognized as an additional category of alluvial deposit; they could indicate either an abundant supply of material, steep slopes, or a climate where the rainfall is either seasonal or irregular (Nilsen, 1982).

It has since been recognised, however, that in nature there is a complete gradation in channel morphology between the straight, braided, and meandering end-members. Lateral accretion, once thought to be diagnostic of meandering rivers, has now been documented in braided rivers

(*eg.* Allen, 1983); high-sinuosity meandering streams can contain bars or islands and display many characteristics (for example Sp) once thought typical of braided rivers (Jackson, 1978; Plint, 1983). As a result, interpretations based solely on the interpretation of a single vertical profile could misrepresent the geometry and internal structure of large bar and channel deposits (see Miall, 1984). A complete spectrum of sediment gravity flows has also been recently recognised (Lawson, 1982; Pierson, 1981; Nemec & Steel, 1984) which range from fairly slow moving high strength/ high viscosity debris flows to swift fully turbulent fluidal flows (hyperconentrated flood flows of Smith, 1986; or sheetfloods). The latter are transitional to heavily sediment laden stream flows; Beverage and Culbertson (1964), for example, have described high suspended sediment concentrations within several streams in the semi-arid southwestern USA which are 'too thin to plow and too thick to drink'.

Full appreciation of the nature and controls of channel morphology can only be attained by consideration of both lateral and vertical variation in lithofacies(see Miall, 1985; Allen, 1983). In this study, channels are divided for purposes of discussion into fixed (ribbon-shape geometry), mobile (broad and shallow with complex fill geometry; braided or meandering types), or sheet-like (essentially unchannelised) following Friend *et al* (1979) and Friend (1983). Vertical sequences from each type of fluvial channel will be examined in tandem with analysis of associated alluvial plain environments for each depositional system. Controls on styles of fluvial sedimentation, as well as the temporal and spatial variation within alluvial sequence, will be discussed later (Chapt. 4.4.2)

4.2.3 DELTAIC COMPONENT

In modern coarse-grained deltas, the accessibility of delta plain deposits has resulted in preferential documentation of subaerial components. Several depositional settings have been recognized including

both stream and mass-flow dominated fan deltas (Sneh, 1979; Wescott & Etheridge, 1980; Schramm, 1981), coastal outwash plains (Boothroyd & Ashley, 1975; Hine & Boothroyd, 1978), bayhead deltas (Syvitski & Farrow, 1983; Prior *et al*, 1986) and short headed stream deltas (Flores, 1975).

A corresponding spectrum of delta types has not been recognized in ancient successions. The subaerial component, although an integral part of coarse grained deltaic systems (Moore & Asquith, 1971; McGowen & Scott, 1974; Sykes & Brand, 1976) is not always well preserved, or simply proves difficult to interpret. Detailed studies have concentrated mainly on processes of sedimentation on the delta front (*eg.* Nemec *et al*, 1984; Pollard *et al*, 1982; Farquharson, 1982; Van der Meulen, 1983; Massari, 1984; Postma, 1984; Surlyk, 1984). Comparative studies have related variation in the nature of sediment dispersal and eventual deposition to aspects of the receiving basin (*eg.* Wescott & Etheridge, 1980; Etheridge & Wescott, 1984), rather than the nature of a particular alluvial system. Considerations of the subaerial component commonly fail to discriminate between fluvial systems of alluvial fans and those of braidplains (Rust,1979). As a result, all braided streams building out into standing bodies of water were treated as fan deltas, even where deposit morphology and facies distribution suggest otherwise (*eg.* Hobday, 1974; Galloway, 1976; Boothroyd & Nummedal, 1978; Harbaugh & Dickinson, 1981; Galloway & Hobday, 1983, p.40; Kingsley, 1984; Pudsey, 1984). In other cases, a fan morphology is assumed from isolated vertical sections (*eg.* Belt & Bussieres, 1981; Feary, 1984; Bergh & Torske, 1986; Flint *et al*, 1986). This overextension and misapplication of the term fan-delta to all coarse grained delta sequences has been noted by several authors (Sykes & Brand, 1976; Dixon, 1979; McPherson *et al*, 1987, 1988; Orton, 1988-enclosure 4).

A fan delta, *sensu stricto,* represents progradation of an alluvial fan into a standing body of water (definition by Holmes, 1965; McGowen,

1971). In ancient sequences, alluvial fans can be distinguished from braidplains on the basis of depositional process: likelihood of debris flows and highly concentrated sheetfloods, radial palaeocurrents, rapid lateral and downfan variation in grain size and attendant facies and a transverse relationship to the structural grain or configuration of a basin (Bull, 1972; Rust, 1979). In contrast, **braidplain** deltas, as defined here, are associated with broad systems of low sinuosity braided streams which bear no strict affinity to a single alluvial fan. Modern examples include coastal outwash plains in Alaska or Iceland (*eg.* Hine & Boothroyd, 1978). The term braid delta has been proposed (McPherson *et al*, 1987), but the term braidplain delta is preferred here to emphasize the deposit morphology as well as the nature of the contributary alluvial system. It is suggested that the term "braid" delta be restricted to deltas whose braided distributary systems are merely derivative from the progradation of solitary braided or low sinuosity rivers; recent examples include bayhead deltas in the fjords in Western Canada, short headed stream deltas on Lake Erie (Flores & Ferms, 1970), or the Copper river delta (Galloway, 1976). The distinction between "braid deltas" and "braidplain deltas", as defined here, is important as it has direct bearing on the difference in alluvial architecture, including the geometry, of resultant sandstone bodies.

In this chapter, 8 coarse-grained deltaic depositional systems are reconstructed using a network of closely spaced vertical sections with lateral tracing of lithofacies in areas of good exposure. Emphasis is placed on description and analysis of distal alluvial plain and nearshore delta front deposits, where fluvial and marine interactions result in delicate and complex environmental differentiation. Fan deltas (4 examples) are distinguished from braidplain deltas (3 examples) and a 'common' delta based on the geometry and channel type of the contributing alluvial input. The latter two types of delta are basin axial systems fed

laterally by alluvial fans. The alluvial component of the coarse-grained deltaic systems will be considered first in the discussion (Chapt 4.4.2). It will be later be demonstrated (Chapt. 4.4.3) that both the nature (fixed or mobile channels, unchannelled) and the large scale geometry (alluvial fan or braidplain) of the fluvial input controlled the efficiency of discharge in coping with basin energy, the patterns of sediment dispersal during wave reworking, and the type of associated coastline.

4.3 DEPOSITIONAL SEQUENCES

4.3.1 GWERN GOF DEPOSITIONAL SYSTEM: A LARGE FAN DELTA?

Sediments subjacent to the Gwern Gof Tuff in the Tryfan anticline comprise two large scale (60 m) coarsening upward cycles (A & B) each interpreted to represent an episode of delta progradation, separated and overlain by thin delta abandonment horizons (Fig. 4.3). Only the lower cycle (A) will be described in detail (depositional phases 1-6 in Fig. 4.3).

4.3.1.1 THE DELTA FRONT

The lower coarsening upwards sequence (Fig. 4.3, sequence C2, phase 2) is made up of thin (10-20 m) discrete packages of fine-grained sandstone which are thought to represent deposition of channel-effluent derived deposits (distributary mouth bars). Their thin size reflects an inconstant sediment supply due to frequent channel avulsion. The inferred distal mouth bar (Fig. 4.3, sequence C2, phase 2) is dominated by graded beds up to 20 cm thick. A subtle upward gradation from very fine-grained to fine-grained sandstone is accompanied by thinning of beds to 10 cm in the proximal mouth bar. The beds probably reflect density currents from individual river-flood events. The second coarsening upward deltaic sequence (Phase 7) indicates a higher, more rapid supply of sediment into the nearshore zone. Distal mouth bar deposits are still pervasively bioturbated (Fig. 4.4) but the more

proximal sandstones display distinct horizontal bedding (Fig. 4.5). Occasional wave ripples and thin units (10cm) of sigmoidally cross-stratified sandstone (?tidal currents) indicates minor basinal influence.

4.3.1.2 A DISTRIBUTARY CHANNEL

A channelised fining upward unit of cross-stratified medium sandstone (Fig. 4.3, Section C2, Phase 4), overlying the first delta front sequence is interpreted as infill of a broad low sinuosity distributary channel. Scour relief indicates that the channel cut at least 4.5m into a thin (less than 8m) laterally continuous (extent more than 500m) fine grained interval which coarsens upwards from mudstone to coarse siltstone. This subjacent unit probably represents a broad interdistributary bay, with occasional thin (15cm) graded or horizontally laminated beds of coarser material representing overbank flooding (Elliott, 1974). The channel itself was at least 8m deep (infill thickness) and 150m wide (outcrop availability). Its infill gradually fines upwards from comparatively well sorted, texturally mature medium- to fine-grained sandstones and typically displays a consistent vertical sequence of lithofacies (Fig. 4.8). Tabular planar cross statification (phase 4B in Fig. 4.8) with tangential to angular foreset strata and unidirectional dip aximuths to the west is the dominant sedimentary structure (Facies Sp, Spt; Fig. 4.9, 4.10). Foresets commonly display "hooked" bases (Fig. 4.9) which suggests some unsteadiness in discharge. Vertical stacking of cross-sets suggest that the channel was filled by rapid aggradation on coalesced transverse bars or "sand flats" (*eg.* Smith, 1972) rather than by lateral accretion. Migration of the channel to form a braidplain was inhibited by the cohesiveness of muddy substrate and banks (*cf.* Galloway, 1981).

4.3.1.3 A BRAIDED STREAM

Pebbly sandstones within more proximal alluvial deposits capping the first coarsening-upward cycle of the Gwern Gof Mbr. (Fig. 4.3, sequence C4,

phase 4) display two contrasting facies assembages, the lowermost of which represents an "ideal" sequence of lithofacies which one would expect from infill of a low sinuosity braided stream (Fig. 4.11).

Four channel fill units, which vary in thickness from 50 to 500 cm, can be recognised. The base (where exposed) of all sandstone units is sharply erosional with high (50-100 cm) local relief. Internal sedimentary structures (St, Sp, Sh) display a consistent vertical organisation along with a gradual upwards fining from medium-coarse to fine sandstone. The basal fining-upward sequence displays a complete tripartite sequence with trough and then planar tabular cross-stratification overlain by horizontally laminated fine sandstone with occasional small scale (1cm) ripple cross-lamination. The upper channel fill units are thinner, with less preservation of both planar cross-stratification and horizontal lamination. Trough cross-stratification is the dominant lithofacies and commonly shows alternation of coarse and very fine sand along foresets.

In light of their stratigraphic position, the fining-upwards sequences with their unidirectional palaeocurrents, clearly represent deposits from fluvial channels. The grain size, the -ratio of sandstone to siltstone, and most importantly the abundance of planar cross-stratification within lower to mid-portions of the sequences, all suggest deposition within low sinuosity sandy braided streams (Jackson, 1976; Miall, 1977; Cant & Walker, 1976, 1978). Trough cross-stratification (Fig. 4.7) probably reflects migration of sinuous megaripples within the base of channels, whereas the planar cross-stratification is attributed to high stage deposition within transverse or linguoid bars (Smith, 1972). The capping lithofacies of horizontally laminated sandstone, and small current ripples, implies deposition by either transitional to upper flow regime plane bed, or migration of lower flow regime low-amplitude bedforms (Smith, 1971), during falling water stages on a shallowing (*ie.* 5 cm deep) channel floor.

The comparative absence of planar cross-stratification within the upper channel fill sequences probably stems from a combination of (1) shallower flow depths which prevented development of channel bars (see Chapt. 4.3.2.3), or (2) more pronounced lateral migration of dune-filled channels which a) removed any evidence for existence of transverse bars (*cf.* Campbell, 1976) and b) removed the top portions of underlying sequences. If one assumes comparable velocities between distributary channel and braided stream portions of the fluvial system a channel width of at least 300 metres (4 m deep channels) is needed on the braidplain to carry the same volume of discharge (*ie.* same cross-sectional area) as the distributary channel (150 m wide x 8 m deep).

The top portion of the alluvial sequence (Fig 4.11), indicates an abrupt change in fluvial style. It consists largely of horizontally bedded medium-coarse sandstone intercalated with siltstone. The sandstone units range in thickness from 4 to 50 cm, are internally structureless, and contain angular rip-up clasts of fine-grained rhyolitic ash-fall tuff (up to 8 cm long). They are interpreted as thin debris flows or sheetfloods. Cross-stratification exhibits extremely low foreset angles which reflect high flow velocities (*cf.* Jopling, 1965). The only channel developed was filled after its abandonment by alternating slurries of medium sandstone and siltstone. The above assemblage of lithofacies indicates extreme fluctuation in both the amount of fluvial discharge and sediment water ratio. Distinct units of primary rhyolitic air fall tuff can be recognised within adjacent siltstones; they indicate that this fluvial transformation, and cessation of sediment supply to the first delta lobe, was directly associated with a brief episode of volcanicity in the source area. The occurrence of debris flows in the capping facies association, the radial pattern of fluvial palaeocurrents (*cf.* sequences C2 & C4 in Fig. 4.3), and a transverse relationship to the basin margin suggest that the fluvial

system was associated with a large alluvial fan, probably reflecting uplift on the Conwy fault, 10 km to the east (Fig. 4.12). The abundant supply of mud, derived by chemical weathering of subjacent volcanics, forced the distributary braided streams to form deeper channels on the lower delta-plain. Galloway (1981) has described similar, bedload streams of the Cenozoic Gueydan fluvial system which were confined due to deposition of cohesive, rapidly altered volcanic ash.

4.3.1.4 DELTA ABANDONMENT AND PALAEOGEOGRAPHY

Delta abandonment facies vary according to position on the delta front. Bioturbated fine sandstones (Fig. 4.8; phase 5A) represent transgressive remnants of distributary channel deposits. The overlying mudstone (Phase 5B) extends laterally for 1500m. It contains rare thin sandstone beds and is interpreted as representing a back barrier lagoon. A suprajacent 5m thick sequence of well sorted, magnetite-rich fine to medium sandstone containing moderate scale (20cm) trough cross-stratification and horizontal lamination (Fig. 4.3, sequence C2, phase 6; see Appendix 5) represents progradation of a 'Chandeleur-type' (Elliott, 1986, p.140) barrier island. Shoreface lithofacies and heavy mineral concentration indicates significant nearshore wave energy in the absence of fluvial input. Similar sequences have been described from modern non-barred high energy coastlines in California and Oregon (Clifton *et al*, 1971; Howard & Reineck, 1981).

Interpreted palaeogeography during maximum progradation of the fan delta indicates a lobate geometry of the delta front and a moderate extent of basinward sediment transport (Fig. 4.12). The continuous delta front sand sheet resulted from lateral coalescence of the lobate "unit" mouth-bars of closely spaced distributary channels, a pattern of sedimentation similar to that of shoal water lobes of the modern Mississippi delta (Elliott, 1986). As the delta facies are dominantly fluvial or effluent-

derived, despite the considerably high wave-energy levels inferred, the Gwern Gof fan-delta is categorised as fluvially dominated variety.

4.3.2 TRYFAN FACH DEPOSITIONAL SYSTEM

4.3.2.1 INTRODUCTION TO BRAIDPLAIN DELTAS

Deposition within braidplain deltas was the dominant mode of basin infill within the Tryfan Formation. These depositional systems prograded toward the south and southeast and were fed laterally by tributary alluvial fans at Craig yr Ysfa, reflecting uplift both along the Conwy or related subparallel fault systems, and to the SW of Llanberis pass (inferred). Variation in sediment composition (rhyolitic pebbles, magnetic heavy minerals etc.), airfall tuffs and stratigraphic position allow correlation of depositional packages on the alluvial fan sequences with various braidplain sequences to the southwest. Three of these braidplain deltas, the Tryfan Fach, Capel Curig, and Magnetite Mbrs., including the contributing alluvial fan at Craig yr Ysfa will be described in this chapter, with the sedimentology of the stratigraphically youngest Moel Siabod Mbr. discussed later (Chapt. 6.2.2) in relation to the overlying 1st Mbr. of the Capel Curig Volcanic Formation.

4.3.2.2 TRYFAN FACH DELTA FRONT: A MUDDY COASTLINE

The delta front of the Tryfan Fach braidplain delta (Fig. 4.13) is dominated by fine grained lithologies and interpreted to represent progradation of a barred muddy shoreface. Occasional brachiopods (*Rostricellua*) indicate a marine setting. Sediments are organised into several coarsening upwards bipartite units 5-15m thick. Basal portions (*eg.* phase 1), consist of bioturbate mudstone with storm generated massive or horizontally laminated beds of coarse siltstone and sandstone (*cf.* Brenchley, 1985). The upper portions of sequences are thought to represent both sand bars and shoals which vary according to position relative to the postulated shoreline. Medium to coarse sandstone within offshore bars

(sequence C9, phase 2) contain large scale(25cm) trough cross-bedding indicating deposition under considerable wave energy. The equivalent nearshore shoals (sequences C5B, C7, C1, C6; phase 2 & 3) consisting of rippled coarse siltstone with occasional undulating lamination, and small scale (10cm) trough cross-strata indicates a pronounced reduction in wave energy toward the coastline. As the above depositional packages can be correlated over considerable distances along depositional strike, the nearshore shoals themselves probably had a linear morphology. The variety of internal sedimentary structures would reflect subtle variation in both grain size and wave energy; well-described sandy shoals from a Carboniferous wave influenced shallow marine platform (de Raaf *et al* , 1977) display a similar spectrum of lithofacies.

It is envisaged that progradation of chenier plains in Australia (Rhodes, 1982) or muddy coastlines in Surinam (Wells & Coleman, 1981) could produce the above patterns of delta front sedimentation. In these situations, mud can easily accumulate in environments of at least moderate wave energy especially when the sediment supply is large; soft muddy bottoms and subaqueous fluid mud-shoals are, in fact, extremely efficient at completely attenuating shallow water wave energy, protecting shorelines from erosion and providing conditions favourable for further sedimentation. Development of sand bars is related to periods of wave reworking during reduced mud input. The thin (2-4m) unit of bioturbate siliceous mudstone (Fig. 4.15) with occasional wave or current ripples capping the delta front in these sequences is thought to represent the protected shoreline itself. Walker & Harms (1971) have described a similar coastline from the muddy Upper Devonian Catskill complex of NE USA.

4.3.2.3 TRYFAN FACH BRAIDPLAIN: A PRODUCT OF VOLCANICITY

Fluvial deposits of the Tryfan Fach braidplain consist of poorly sorted very coarse to very fine sandstones (Fig. 4.14) which form a

multistorey laterally extensive (greater than 1.5km) sheet directly overlying the muddy coastline sediments. Tuffaceous siltstone rip up clasts and subangular rhyolitic pebbles are common. The sheet sandstones were deposited as broad tabular units 1-3m thick separated by thin beds of siliceous siltstone (Fig. 4.14) which are laterally continuous for only short distances (200m) on outcrop scale. A notable marker horizon of particularly tuffaceous "sea green" siltstone, however, can be traced for at least 900 m. Sandstone units are generally characterised by extremely rapid vertical changes in grain size and texture and usually rapidly fine upwards from very coarse sandstone to siltstone. These minor sequences occasionally display basal scours. Internal sedimentary structures often show a consistent vertical sequence (Fig. 4.16). Trough cross-bedding is the dominant lithofacies with subordinate planar crossbedding or horizontal lamination. Set size of crossbedding decreases upwards within cycles (25cm to 5cm) in conjunction with grain size changes. Upper portions of sequences consist of fine sandstone to siltstone with horizontal lamination or ripple cross-lamination. The siliceous composition of these siltstones, and rare occurences of accretionary lapilli provide evidence for contemporaneous rhyolitic volcanicity (Fig. 4.17).

Geometry of the above sandstone packages, internal structures and coarse texture indicate deposition within a hierarchy of small braided channels. The lateral extent of channel fill sequences indicates that maximum width of major channels was 200m. The prevalence of trough cross-bedding and relative absence of planar cross-bedding suggest that channel bars were insignificant due to either shallow depths of flow and/or removal by channel migration. Lateral variation in the relative thickness of interpreted fluvial and floodbasin components along depositional strike indicate that the most persistent axis of coarse fluvial deposition and sand transport was near the SW side of the braidplain close to sequence C7.

More detailed stratigraphic correlation on the braidplain, however, other than that provided by the distinctive green siltstone horizon, is difficult. In the NE (section C6), successions indicate infill of several 3-4 m deep channels (Fig. 4.14). Fluvial discharge and attendant sediment/water ratios appear to have been more variable along the SW side of the braidplain. In section C5B, depositional sequences indicate that thin debris flows, and sheetfloods were an important component of deposition on the alluvial plain. The thickness of channel fill sequences at this locality suggests that initially most channels were less than 1 m deep, with more substantial fluvial systems only developing later.

The contrasting styles of fluvial transport between opposing sides of the Tryfan Fach braidplain are probably a manifestation of more direct input of volcanically produced (or induced) sediment loads onto the SW portion of the braidplain (Fig. 4.19). Petrographic evidence (Chapt 3.4.2; Fig. 3.111 and 3.128) has already been cited for penecontemporaneous pyroclastic activity of rhyolitic composition in this direction. Associated tectonic must be invoked to account for the stepwise lateral migration of the main fluvial channel to the SW across the braidplain.

Fluvial deposits in all the braidplain sequences are gradationally overlain by a finer-grained interval of interbedded sandstone and tuffaceous siltstone (Fig. 4.13). Sandstone units vary substantially in composition and thickness. Very coarse sandstone (beds 20-50cm thick) usually contains small scale (10 cm) trough cross-stratification which yields palaeocurrents toward the SE. Climbing ripple cross-lamination and fine horizontal lamination are the dominant lithofacies within the finer-grained and thinner (10-15 cm) sandstone beds. The intervening siliceous siltstones display horizontal or undulating lamination and small scale rippling. The proportion of sandstone to siltstone decreases upward through the sequence. Sub-horizontal burrows (*Cruziana* ichnofacies: *Rhizocorallium;*

Fig. 4.18) indicate a low energy depositional environment (Frey & Pemberton, 1984, p.193), but cannot be used to infer water depths. *Rhizocorallium* has also been described from coastal lagoon or freshwater lacustrine settings (Fürsich & Mayer, 1981; Ekdale *et al*, 1984, p.162). The above facies association is interpreted as representing a transgressive coastal plain to lacustrine setting, and reflects a gradual waning of sediment supply to the braidplain. Similar lithofacies associations have been described from floodbasin sequences of the Hornelen basin (Steel & Aasheim, 1978). The sandstone units either reflect infill of small ?low sinuosity channels or deposition by sheetfloods into shallow (1-2m deep) water. Some of the thin siltstone units obviously represent primary pyroclastic deposits; most, however, likely represent redeposition of the fine siliceous material by muddy low density sheetfloods.

Sediment composition within braidplain and delta front settings all indicates progradation in conjunction with a period of active rhyolitic volcanicity (Fig. 4.19). An initial abundant supply of fine grained silts and mudstones was reworked by waves of moderate energy into a series of shore parallel shoals and bars, which attenuated wave energy enabling a muddy shoreline to prograde. Sediment supply to this coastline was likely by small channels or sheetfloods (*cf*. Wells, 1983). With a sudden increase in sediment grade and abundance, a transformation of fluvial style resulted in a brief period of rapid aggradation both on tributary fans and within the broad braidplain. This braidplain prograded rapidly as a sheet across the preexisting shoreline and met little resistance from wave attack. The Tryfan Fach successions thus represents a fluvially dominated braidplain delta.

4.3.2.4 TRYFAN FACH ALLUVIAL FAN: A VOLCANIC BYPRODUCT

The lowermost coarsening upward sequence at Craig yr Ysfa (Fig. 4.20, 4.21) represents one of the dominant contributing alluvial fans for the

Tryfan Fach braidplain. The base of the sequence (Phase 1) consists of bioturbated siltstones to very fine sandstones which are thought to reflect deposition in shallow ?lacustrine environments. These sediments are overlain by a thin sequence (Phase 2-4, Fig. 4.21) consisting of thin beds(10-50 cm) of horizontally laminated very-fine sandstone. Silty flasers occasionally define straight-crested current ripples (wavelength 5 cm), intercalated units (5 cm thick) of primary rhyolitic airfall tuff indicate concurrent volcanicity. Based on its stratigraphic location, this facies association must record the transition from lacustrine to fluvial environments. The horizontally laminated sandstones suggest upper flow regime conditions and could either represent sheetfloods into low wave energy lacustrine environments or wave-swash on a low energy shoreface. The preservation of the airfall tuff horizons, and palaeocurrents toward the inferred offshore (WSW) favour the former interpretation.

Alluvial fan sediments erosively scour 4.5 m into the subjacent sheetflood deposits over a horizontal distance of 50 m. The basal fining-upwards sequence (phase 5A) of medium coarse pebbly sandstone would thus represent infill of a low sinuosity distributary channel by lateral migration and aggradation of sinuous crested dunes or megaripples. Suprajacent distal to mid-fan sediments consist of either (1) poorly sorted medium sandstone with trough cross-bedding (Phase 5C, 5F) or (2) channelised, normally graded, conglomeratic sandstone to fine sandstone (Phase 5B, 5D). The former lithofacies would reflect deposition within shallow (« 1 m deep) ?braided streams; the latter lithology indicates increased sediment/water ratios and deposition by fluidal sediment gravity flows. The mass flows are either internally structureless or contain poorly developed trough cross-stratification within the finer-grained upper portions. The latter relationship indicates an initial rapid fallout of suspended load and deposition by direct suspension sedimentation, followed

transitionally by traction sedimention within the more dilute low-density upper portions (see Lowe, 1982). The small-scale intercalation of debris flows with stream flows of contrasting grain sizes indicates rapid variation in fluvial discharge conditions.

More proximal alluvial fan sediments (Phase 5E, 5G) are dominated by unstratified conglomeratic sandstone occurring in tabular beds 20-50cm thick. They commonly contain subvertical rhyolite pebbles (4cm size) or rip up clasts of siliceous siltstone (30cm long). The sharp non erosive bases, occurrence of well defined units, lack of sorting or grading and nature of fabric are all in accord with deposition by cohesive matrix supported debris flows (Lowe, 1979; Rust, 1979). A thin isolated lense of imbricate clast-supported conglomerate probably represents a gravel bar (*cf.* Ramos & Sopena, 1983); the associated wedge of low angle cross-stratified sandstone which is sub-parallel to the margins of the conglomerate lense would then represent lateral accretion during waning flow and channel abandonment.

The overall vertical organisation of the alluvial megasequence implies rapid progradation of a small flashy alluvial fan westwards into the Tryfan depocentre. The contrast in petrographic composition between these alluvial fan sequences and synchronous braidplain sequences in the Tryfan anticline has already been noted (Chapt. 3.4.2, Fig. 3.111, 3.128), with sediments at Craig yr Ysfa derived by erosion of a mixed tuffaceous (rhyolitic) sediment/rhyolite intrusion/granite source. One cannot thus appeal to a direct volcanic control on progradation of this particular alluvial fan, as is possible with the braidplain sequences. It is thought, however, that tectonic changes associated with the concurrent volcanicity, as expressed in this case by the thin units of airfall tuff subjacent to the main episode of coarse clastic deposition, produced enough topographic relief between Craig Yr ysfa and Tryfan Fach to initiate progradation of the entire coarse-grained alluvial fan at Craig yr Ysfa into the Tryfan

depocentre. This alluvial fan can then be viewed as a secondary by-product of the volcanicity which dominated concurrent progradation of the braidplain sequences to the south.

Sequences from abandonment of this alluvial fan (Fig. 4.21, 50-60 m) are extremely thin and fine upwards from very coarse pebbly sandstone to fine sandstone, with a corresponding change in facies from sediment gravity flows, through trough cross-stratification to horizontal lamination. Reasons for fan cyclicity and abandonment of this and other alluvial fans will be discussed later (Chapt. 4.4.2.2, 6.3).

4.3.3 MAGNETITE MBR.: CONCURRENT BASALTIC VOLCANICITY?

The Magnetite Mbr. of the Tryfan Fm. comprises a thin horizon of poorly sorted very coarse sandstone to mudstone which contains abnormally high concentrations of magnetitic heavy minerals (Fig. 3.71, 3.72, 3.110, 3.111) and is in fact responsible for one of the dominant local aeromagnetic anomalies in North Wales (R. Evans, BGS Keyworth, pers. comm.). Sediments on the west side of the Tryfan anticline are organised into a poorly defined coarsening upward sequence (Fig. 4.22) which, in a fashion similar to the Tryfan Fach depositional system, is interpreted to represent progradation of a braidplain delta.

4.3.3.1 THE MAGNETITE MBR. DELTA FRONT

Sequences interpreted as representing the delta front (Fig. 4.22) can be divided into two stratigraphic packages. "Prodelta" deposits (phase 1) are dominated by uninformative cleaved mudstones. Sandstones units are all less than 1 metre in thickness; small sets of contained trough cross-stratification yield palaeocurrents with variable dip azimuths. They are interpreted to represent small sandstone bars in a shallow ?lacustrine environment. The upper portion of the delta front is dominated by fine-medium grained sandstone (phase 2) which display a comparatively wide variety of sedimentary structures. A thin unit of well sorted trough

cross-stratified medium sandstone which yields palaeocurrents parallel to the inferred shoreline, probably represents some sort of nearshore sand bar. The overlying sandstones contain undulating horizontal lamination, ripple cross-stratification, planar low angle (5° dip) lamination, and isolated sets of low angle (20° dip) tangentially-based cross-stratification. The lithofacies association is thought to represent progradation of a sandy shoreface. The small sets of tabular cross-stratification yield palaeocurrents similar to the overlying fluvial cross-strata and probably represent the initial incursions of fluvial flood events. The only possible evidence for any basin energy is provided by the interstratified units of low-angle cross-stratification and undulating lamination which might represent brief periods of wave reworking.

4.3.3.2 FLUVIAL SEQUENCES

The petrography of these fluvial sandstones (Fig. 3.110, 3.111) indicates derivation from both rhyolitic and basaltic sources, with the proportion of basaltic detritus increasing upwards. Sandstones display considerable local variation in texture and composition, as illustrated by a small half-graben (NE-SW extension) infilled by granule conglomerate (Fig. 3.110). Local occurrences of convolute lamination (Fig. 4.24) reflect plastic deformation of partially liquefied sediment soon after deposition.

Trough cross-stratification is the dominant sedimentary structure (Fig. 3.71, 4.25) with individual sets showing significant local variation in grain size and attendant composition. Coarse-grained quartzo-feldspathic material with rhyolite pebbles characterizes the bottom of foresets with fine-grained magnetite-rich chloritic detritus along the upper portions. The trough cross stratification is often organised into small cycles (less than 1 m) with set size and grain size both decreasing upwards, and compositional changes similar to those observed on a smaller scale within individual sets. The above cycles are interpreted to represent infill of

small low-sinuosity streams. Convincing evidence for chanelling is only seen at the base of the lowermost fining-upward cycle which cuts 70 cm into the subjacent shoreface deposits. Other channel fill deposits have tabular geometries. Major cycles extend as lateral (transverse to palaeocurrents) sheets for distances greater than 250 m and indicate widespread migration of channel systems across the braidplain.

Magnetite heavy minerals often occur as thin laminae (up to 2 cm thick) along the tops of foresets (*ie.* where flow separation took place) and often merge upwards as the tops of fining-upwards cycles into 10-15 cm thick zones of horizontally laminated magnetite rich sandstone interpreted to reflect upper flow regime plane bed conditions (Fig. 3.72). The association of these two sedimentary structures is thought to be genetic with the complete fluvial cycle (Ss - St - Sh) and transition from lower to upper flow regime conditions reflecting both shallowing of channel due to its infill and a decrease in discharge capacity relative to sediment load. Minor occurrences of low angle cross-stratification (Fig. 4.23) which dips transverse to the dominant palaeocurrent direction may reflect lateral infill of abandoned scours.

The rapid cessation of sediment supply to the braidplain was associated with deposition of a sheet (Fig. 4.22, 4.26) of thin (10 cm) pebble conglomerate beds separated by diffuse zones with horizontal magnetite laminae (Fig. 4.26).

The origin of fluvial sequences from the Magnetite Mbr. is enigmatic:

(1) The variation in sandstone texture and composition, preservation of magnetite on the tops of foresets or "peppered" throughout sediment, and presence of soft sediment deformation all indicate rapid rates of sedimentation. The frequent changes in style of fluvial sedimentation with sediment transported either as sheetfloods or as bedload within braided fluvial channels points to repetitive fluctuation in the sediment/water

ratio (?sediment load) supplied to the depositional system.

(2) The proportion of heavy mineral grains within the fluvial sequences, however, suggests several cycles of erosion and deposition are necessary to account for their observed concentrations. Magnetite and ilmenite alone comprise on average 20% of the detrital framework within sandstones; this would entail a 2-fold concentration of the highest opaque heavy mineral suite typically found within tholeitic basalts (K. Cox, pers. comm., 1988).

The latter situation seems to demand sudden access of depositional systems to a previously "hidden" source of weathered basalt. It is easier to envisage the abrupt arrival of a largely undiluted package of basaltic detritus from redeposition of ash produced by concurrent volcanic activity of this composition. It is thought that first cycle fluvial transport could provide the mineral concentrations necessary by hydrodynamic separation of the bedload and suspended load grain fractions. Most of the detrital magnetite, ilmenite and plagioclase crystals would travel as bedload and form the fluvial deposits studied; the fine-grained chloritic mud resulting from alteration of basaltic glass or weathering of the matrix component of basaltic lava would be transported as suspended load to form the fine-grained delta front (presently chloritic slate) of the braidplain delta.

A possible reconstruction of palaeogeography during maximum progradation of the Magnetite Mbr. is shown in Figure 4.27. A NW-SE fracture passing near Glyder Fawr (SH_6450_5825; hereafter termed Glyder Fawr fault) is suggested as a source for the basaltic pyroclastics. The existence of a zone of structural weakness with this trend (probably a splay of the Llanberis Pass Fault) at this locality, is indicated by a later sheeted dike complex which cuts through the Upper Cwm Eigiau Fm. The low wave energy of the lacustrine environment allowed uncontested advancement of the volcaniclastics within fluvial systems southwards into

the basin; the Magnetite Mbr. would thus be classified as a fluvially-dominated braidplain delta. A thin horizon (section C9, 47-50 m) of bioturbate chloritic and magnetitic very fine sandstones on the east side of the Tryfan anticline would represent the distal portion of the delta front.

The relationship between the above sequences in the Tryfan anticline and correlative alluvial successions at Craig yr Ysfa (Fig. 4.22) is less certain. The grain size and thickness of preserved sediments suggests that there was little uplift on bounding faults along this side of the basin during deposition of the Magnetite Mbr.. The magnetite rich sandstones at Craig yr Ysfa could represent either a distal portion of braidplain sequences to the south, or detritus from an additional coeval volcanic vent (Fig. 4.27).

4.3.4 CAPEL CURIG DEPOSITIONAL SYSTEM.: VOLCANIC ASSISTANCE.

The Capel Curig braidplain delta (Fig. 4.28) represents progradation, strong aggradation, and eventual retreat of a fairly persistent depositional system. Similar to its older cousins in the Tryfan anticline, it was fed laterally by alluvial fans reflecting uplift along the Conwy (Fig. 4.20) and Llanberis pass fault systems. Statigraphic correlations within braidplain sequences indicate three main episodes (A to C in text and on logs) of deltaic advance. These display considerable variation in sedimentary structure and depositional sub-environments according to the rate of sediment supply relative to basin energy. Each individual episode of delta progradation is described separately below, with controls on the sedimentation of the entire depositional system discussed later (Chapt. 4.3.4.8).

4.3.4.1 EPISODE A: A BARRED COASTLINE

The first deltaic succession (Fig. 4.28) is interpreted to reflect progradation of a barred shoreface. Its basal portion (Appendix 3, section

C9, 80 - 110 m; abbreviated hereafter as C9, 80-110m) consists of fine siltstone with numerous thin 5-15 cm beds of fine sandstone which increase in both abundance and thickness upwards. Sandstone beds are thought to reflect deposition by underflows from alluvial flood events. Sedimentary structures include horizontal to low angle lamination, undulating lamination and ripple cross-stratification. More proximal sequences (Fig. 4.28, phase 1B; C9, 110-130 m) are dominated by sandstones which either contain horizontal bedding, fine horizontal lamination, or small scale (5-20 cm sets) trough cross-stratification. Concentration of magnetic heavy minerals commonly occurs along the tops of foresets, or within rippled horizons suggesting periods of marine reworking.

Correlative sediments of the NW side of the Tryfan anticline are coarser grained, but contain a comparable suite of sedimentary structures (Fig. 4.28). Two lithofacies associations can be recognized, depending on the amount of intercalated mud (Fig. 4.29):

1) The lower facies assemblage (Phase 1A) consists of interbedded sandstone and mudstoone with the grain size and thickness of sandstone units increasing upwards. Sandstone beds are initially massive, or contain thin horizontal bedding planes, undulating lamination, or small scale (10cm) trough cross-stratification. Palaeocurrents are towards the inferred offshore (S) or alongshore to the west. They are thought to reflect deposition by flood events or wave-induced currents on the delta front. The common occurrence of magnetic heavy minerals in this offshore setting probably reflects erosion and redeposition of uplifted portions of the subjacent Magnetite Mbr. depositional system in fault blocks to the NW.

2) The upper facies association consists entirely of well sorted medium sandstone with occasional occurrences of brachiopods. Trough cross-stratification dominates with sequences displaying little vertical variation in set size. Most palaeocurrents are toward the inferred

offshore.

The complete succession is thought to reflect progradation of a non-barred shoreface. Although the abundance of trough cross-stratification indicates considerable nearshore wave energy, there is no evidence of pronounced storm activity within this part of the basin.

4.3.4.2 EPISODE B: east side of Tryfan anticline

The second episode of deltaic progradation varies in thickness from 40 m to 75m as controlled by the amount of basin subsidence. On the east side of the Tryfan anticline, the delta front sequence (Fig. 4.30; phase 2A, 2B) coarsens upwards from mudstone to medium-grained sandstones and is thought to represent progradation of a distributary mouth-bar complex. Distal (lower) facies are very fine-grained sandstones showing thin (10-15cm) graded beds, ungraded horizontal layering (1-5cm), low angle planar cross-stratification and occasional ripple cross-sets. More proximal (overlying) facies are medium-grained sandstones with undulatory to horizontal stratification (3-10 mm laminae), reflecting higher levels of both effluent energy and marine reworking.

This sequence (70m) is overlain by two thinner, more complex sequences (phase 3,4 in Fig. 4.30; Fig. 4.31) attributed to progradation of a barred shoreface. The sediments are siltstones to medium-grained sandstones with adjacent beds commonly being of varying grain size. Sandstones display medium-scale (40 cm) trough cross-stratification and planar horizontal lamination, both with heavy mineral concentrates, indicative of high nearshore wave-energy. The lack of any consistent vertical sequence among lithofacies in either succession probably reflects numerous episodes of shoreline progradation and retreat.

These deposits are directly overlain by a thin (4m) laterally persistent (greater than 500 m) unit of interbedded very fine grained, horizontally laminated sandstones (10-20 cm beds) and siltstones (Phase 5

in Fig. 4.30). These are interpreted as storm washover sheets in a muddy back barrier lagoon. The subjacent shoreface sequences would then be part of a larger barrier island complex.

4.3.4.3 EPISODE B: west side of Tryfan Anticline

Lithofacies associations within correlative sequences to the NW indicate a totally different suite of depositional environments. A coarsening upward succession of mudstone to bioturbate fine to medium sandstone (Fig. 4.29; phase 2A, 2B), with no evidence for any wave energy, is thought to represent infill of a small protected lake.

The homogeneity of these delta front sediments contrasts with the variation in sandstone texture and grain size displayed by the overlying sandstone succession. Here, sediments consist of poorly sorted medium to coarse-grained sandstones which are broadly organised into 1-3 m thick tabular units, separated by thinner beds of siltstone. Sandstone petrography (sample C17) and siliceous composition of siltstone beds both indicate some volcanic contribution. The sandstone units have sharp, locally erosive bases and contain abundant (up to 30% by volume) angular rip-up clasts of the subjacent fine-grained horizons. Trough cross-stratification is the dominant sedimentary structure.

Although lithofacies and available palaeocurrents suggest some similarity with a sucession previously interpreted to represent progradation of a wave dominated shoreface (Phase 1A, 1B, Fig. 4.29), the variation in sedimentary texture and size of cross-stratification (*ie.* set size varies from 30 cm to 3cm), common presence of siltstone rip-up clasts, and abundance of siltstone horizons all favour a fluvial origin. The sequence is therefore thought to reflect deposition within low sinuosity braided streams on an alluvial plain. Occasional changes in fluvial regime and sediment supply are indicated by: (1) abandoned channels which were later infilled by siltstone and fine sandstone (section C1, 225.4 m) or (2)

the rapid downcurrent change in grain size from coarse sand to mud on a rare 30 cm thick set of planar cross-stratification (section C1, 222 m).

The above portion of the braidplain was rapidly abandoned, with a thin unit (section C1, 227.2 m) of trough cross-stratified very coarse magnetic sandstone the only possible candidate for transgressive wave reworked deposits. The suprajacent mudstone sequence is interpreted to reflect a return to lacustrine conditions. These "lakes" must have been protected from the more extreme wave regime that existed to the south (Chapt 4.3.4.6) by an array of beach ridges and nearshore bars similar to those indicated by facies associations to the SE (*cf.* Fig. 4.31, section C9, Phase 3,4).

4.3.4.4 EPISODE C: TRYFAN ANTICLINE: A BRAIDPLAIN

The third episode of deltaic progradation (Fig. 4.28, 4.32) represents the maximum advance of the Capel Curig depositional system southwards into the basin. Distinctive occurences of airfall tuff or magnetic heavy mineral laminae (Fig. 3.33) allow detailed (5-10 m scale) correlation of depositional sequences across the Tryfan anticline (ca. 2.5 km unfolded).

On the east side of the Tryfan anticline, the sequence comprises a thick succession of moderately well sorted, medium sandstone (Fig. 4.28, phase 6; Fig. 4.32). These sediments were deposited as broad tabular units 4-7m thick separated by thin laterally continuous beds of siltstone. Units have sharp, occasionally erosive bases and are often overlain by a thin zone of horizontally laminated or rippled very fine sandstone. Trough cross-bedding is the dominant lithofacies (Fig. 4.34) and set size usually decreases upward within sandstone packages. Subordinate sedimentary structures include planar crossbedding and horizontal lamination containing magnetite (Fig. 4.33). Lithofacies and deposit geometry suggest deposition within sandy unconfined braided channels. Sandstone units vary in thickness and sedimentary structures according to position on the alluvial plain. Between fluvial axes (sequence C3), the proportion of channel sandstone to

floodbasin siltstone is slightly less and solitary sets of planar crossbedding representing deposition of sandwaves within transverse bars are significantly more common. These bedforms were not preserved near the main fluvial axis (sequence C9) due to repetitive lateral migration of dune filled channels resulting in a sand rich belt of vertically stacked and amalgamated sequences dominated by trough cross-bedding.

Correlative alluvial plain sediments on the west side of the Tryfan anticline (Fig. 4.28, 4.29; phase 6), were deposited in either shallow (1-3 m deep) lacustrine/floodbasin or braided stream environments. The "lakes" were filled by sheetfloods of siliceous siltstone (Fm, Fg) or crystal-rich (feldspathic) tuffaceous coarse-grained sandstones (Sm, Sg); their is no evidence of any wave activity within the basin. During times of reduced sediment input, higher discharge, or coarser sediment calibre, small ?low sinuosity channels developed on the associated alluvial plain. The maximum channel depth (sequence thickness) was less than 1 metre. They were filled by lateral migration and aggradation of sinuous crested dunes. Vertical and lateral variation in facies are similar to changes found within the Tryfan Fach alluvial plain (Chapt. 4.3.2.3)

4.3.4.5 CAPEL CURIG ALLUVIAL FAN AT CRAIG YR YSFA

Similar to the Tryfan Fach delta, the fluvial systems in the Tryfan anticline were supplied laterally with sediment laden discharge from tributary alluvial fans. At Craig yr Ysfa, 5km to the north, correlative sediments (sequence F1, Fig. 4.20, 4.35, 4.36) consist of interbedded conglomerates (maximum pebble size 2 cm), pebbly sandstones, and sandstones, whose lenticular units are 50-200 cm thick and display mainly trough cross-stratification (sets up to 70 cm thick). Palaeocurrents are unidirectional toward the WSW. Conglomerates attributed to debris flows are subordinate, but present. The coarse overall texture, presence of occasional debris flows, internal sedimentary structures and palaeocurrents

suggest deposition in pebbly braided streams on a medial/lower segment of large stream dominated alluvial fan. Several moderate scale coarsening-upward sequences may represent progradation and/or lateral shifting of individual fan lobes. Crude correlation of these sandstone prisms with the various phases of deposition within the braidplain-deltaic sequences of the Capel Curig Mbr. in the Tryfan anticline is suggested by numbering of phases in Figure 4.36. The total thickness of the alluvial fan sequence at Craig yr Ysfa (42 m thick) is insignificant when compared with the deriviative braidplain-delta sequences (up to 200m thick). The condensed alluvial fan sequence suggests repeated uplift of fault blocks along the basin margin, with stream dissection related to fan entrenchment frequently removing mid- to upper fan deposits. It is likely that a number of small alluvial fans merged downslope to form a broad coalescent fan-system (Fig. 4.28; Fig. 4.30, inset map). The dominant direction of sediment transport swung from westerly to southerly as the fluvial system advance further into the basin and adjusted to basin floor topography. The intermediate portion of the resultant basin-axis braidplain is thought to be represented by fluvial sequences in sections C3 and C9 on the SE side of the Tryfan anticline.

4.3.4.6 TERMINUS OF THE BRAIDPLAIN

A thin succession (Fig. 4.37, sequence B4) at Cwm Clorad Isaf (SH 6969 5616), 3km to the southeast, represents coastal termination of this braidplain and displays the complex interaction between fluvial and marine processes. The sequence comprises intercalated sandstone and mudstone beds with the thickness and proportion of sandstone generally increasing upwards. Lithofacies show considerable vertical and lateral variation and allow recognition of 5 moderate scale (ca. 6-8 m) units of sediment (A-E). Each is interpreted to represent a different episode in advance (A-D) and final retreat (E) of the coastline.

A) The first succession (Fig. 3.38) is dominated by stacked beds (50cm -100 cm) of very coarse pebbly sandstones containing trough cross-stratification (20-40 cm sets), large scale (70 cm) sets of tabular planar cross-strata with tangential foresets, or undulatory low-moderate angle lamination. Several thin cobble layers contained within these coarser beds can be traced horizontally for distances of 20 m (Fig. 4.39). Thinner beds of very fine sandstone with undulatory horizontal lamination or bioturbate siltstone often drape large wave ripples developed on the subjacent units of coarse sandstone (Fig. 4.40; see below).

B) Fine-grained horizons up to 1m thick comprise 30% of the second (B) episode of shoreline advance, and enable recognition of three small-scale (1.5 to 3m) tripartite coarsening-upwards sequences. Mudstone at the base of these cycles contain thin (10-20 cm thickness), sharp-based, beds of massive very-coarse sandstone which can be traced laterally for distances of 5 - 20 m before they either pinch out or are erosively removed. Large wave ripples are often contained on their top surface. The middle portion of cycles consists of medium sandstone with undulatory horizontal to low-angle lamination; in the first cycle, it infills a high-angle ?scour (Fig. 4.41) which laterally removes all evidence (except siltstone rip-up clasts) of the subjacent 30 cm thick mudstone bed further to the south (*ie.* seaward). The upper part of each small scale cycle is characterized by very coarse to conglomeratic sandstone which can display high angle trough cross-stratification in top portions.

C) The third (C) shoreline succession (Fig. 4.37, 4.38) is characterised by stacked 1 m thick beds of fine-medium grained sandstone which initially display undulatory or planar seaward dipping (to S) laminae. Amalgamated sets of hummocky-cross stratification become more common upwards (Fig. 4.42). Sandstone units occasionally contain small pebbles or magnetic heavy minerals along laminae of sedimentary structures;

they are separated by either thin discontinuous (5 m extent) siltstone or cobble horizons. The shallowing-upward succession is partially cut out by a narrow channel (2.5 m deep and 12 m wide) filled by moderate-scale (20-40 cm sets) of trough cross-stratification within medium-coarse pebbly sandstones. Palaeocurrents are unidirectional to the SE. Wave ripples (section B4, 18.3 m) are particularly well developed on the top of a thin (20 cm) unit of coarser-grained sandstone which caps this portion of the succession, and defines a prominent 20 m^2 bedding plane. The ripples are cut out locally on this surface by a narrow (1 m wide), shallow (15 cm deep) channel trending SW (towards 220) which contains scattered well-rounded pebbles (3-7 cm size) of flow-banded rhyolite.

These wave ripples have rounded symmetric to slightly assymmetric profiles with the ripple symmetry index varying from 1.5 to 2.5. They occasionally show internal cross-stratification which usually indicates sediment movement to the north towards the inferred shoreline. In plan view, ripples display a straight to sinuous pattern (Fig. 4.43). Crestlines are traceable for distances of up to 20 m with sinuous types variably being in and out of phase. Termination of crestlines or zig-zag junctions are particularly common. On other bedding planes, ripples often change rapidly from straight crested to more complex discontinuous or linguoid patterns suggesting spatial variation in flow velocity due to the interaction with superimposed unidirectional currents (wave-curent ripples of Allen, 1984, pg. 429). All wave ripples are interpreted to represent shoaling assymetric waves which were occasionally influenced by other nearshore currents (*eg.* edge waves, rip-currents from return of storm water from shoreface).

The common almalgamated sets of hummocky-cross stratification (*ie.* swaley cross-stratification) and undulating low angle lamination within fine-medium grained sandstone beds, and large wave ripples on the tops of

pebbly very coarse grained sandstone beds, all indicate a high storm energy during deposition of the above successions. The horizontally persistent (10's of metres) conglomerate layers (usually a single clast thickness) indicate deposition by an agent which acted uniformly over a broad area (see Bourgeois & Leithold, 1984). They are interpreted to represent wave redistribution of fluvially supplied gravel. The association of the above lithofacies with frequent scour development, and high-angle trough cross-stratification within coarse-grained pebbly lithologies suggests deposition in a nearshore zone (*cf.* Leithold & Bourgeois, 1984; Bourgeois & Leithold, 1984) where interaction of fluvial processes with the above wave regime results in a complex differentiation of sub-environments. Changes in orbital velocity or water depth are not necessary to explain the intimate association of straight crested 2-dimensional ripples and bedforms (*ie.* swaley cross-stratification) commonly attributed to 3-dimensional oscillation ripples (see Chapt. 5.2.1). Flow conditions which are capable of moving very coarse sands as ripples will simultaneously transport finer-grain sizes as hummocky bedforms (Leckie, 1987, pers. comm; see Harms *et al*, 1982, Fig. 2.10 A, B, C).

The low angle planar-undulatory lamination which dips seaward (unit C) is thought to reflect deposition on the upper shoreface or swash slope itself. The preservation of storm events in this depositional setting (see Greenwood & Sherman, 1986) necessitates extremely rapid rates of sedimentation and associated basin subsidence. The thin, massive, normally-graded beds of very coarse pebbly sandstone (*eg.* unit B) are interpreted as flood generated high density turbidites resulting from transformation of fluvial traction currents as they move basinward over the steeper gradient of the delta front (*cf.* Kleinspehn *et al*, 1984). The origin of the thicker (up to 1 m) isolated sandstone beds is variable. Most are thought to have been emplaced by similar sediment gravity flows. They are often massive in

the base, and subvertical clasts of tuffaceous siltstone indicate substantial matrix support (Bull, 1962; Rust & Koster, 1984). Palaeocurrent analysis of the trough cross-stratification in the top portions indicates either onshore(NW), offshore(SE), or alongshore (NE) directions of bed formation. These would reflect migration of lunate megaripples from shoaling waves, and rip or longshore currents respectively (Clifton, 1981). The shallow channel cross-cutting wave ripples on the main bedding plane (Fig. 4.43) and the steep sided scour cutting through the mudstone horizon (Fig. 4.41) are both thought to represent rip channels, and suggest a geographic location within 500 m of the shoreline (Swift *et al*, 1983). In the latter case, the steep channel walls and presence of angular rip-up clasts in adjacent sandstones indicates that the eroded material had substantial cohesiveness. As it is unlikely that the vertically scoured walls could exist at the sediment-water interface for very long, scour formation must be genetically associated with deposition of the overlying fine sandstones (see caption, Fig. 4.41). The large channel cutting the third shoreline sucession (C) would represent erosion and infill of a prograding fluvial distributary (see below).

The remainder of the progradational succession (unit D, Fig. 3.37) is dominated by medium-coarse sandstone which displays less variation in grain size in comparison with earlier episodes (A-C) of shoreline advance. Rare occurrences of large wave ripples (at 19.0 m, 25.4 m) or amalgamated sets of hummocky cross-stratification (Fig. 4.46) record continued storm influence. In general, however, sedimentary structures indicate a stronger control of either fluvial or "fairweather" wave processes on depositional patterns. The greater abundance of heavy minerals within lithofacies also indicates a closer proximity to the swash zone and foreshore where wave processes would concentrate these detrital grains (Reid & Frostick, 1985). Trough cross-stratification with abundant magnetite and ilmenite along

foreset laminae (Fig. 4.45), and unidirectional palaeocurrents to the ESE, for instance, are thought to reflect seaward return of waves from the surf zone (*cf.* Clifton *et al*, 1971; Hunter *et al*, 1979). A solitary set of large scale (1.0 m) planar cross-strata with similar heavy mineral concentrations (Fig. 4.46) can be traced for 20 m along the inferred shoreline (to the NE), and might represent nearshore migration of a sand bar in a longshore trough.

The last event in the progradation of the Capel Curig braidplain delta is well recorded in a 3-dimensional array of blocky crags (Fig, 4.46). Narrow 'scoop-shaped' or low angle channels cut into wave reworked deposits are filled by large-scale solitary sets of low angle (20 to 12° downcurrent) cross-strata. By analogy with earlier channel deposits they are interpreted as representing rapid progradation of a fluvial distributary. Mack (1982) has depicted similar-sized bedforms within alluvial deposits which he attributes to fluvial sheet flow events where flood water emerged from a fan-head. In this case, the discharge from a network of braided streams on the lower delta plain, was probably funnelled through discrete channels as the flow adjusted to shore parallel barrier spits (block diagram in Fig. 4.37). The adjacent wave reworked deposits (Fig. 4.42; 4.46) were probably part of a mouth bar or sandbar which had "welded" across the distributary mouth during a previous period or reduced fluvial input. The actual position through which sediment was delivered from the braidplain would not tend to vary considerably with time; once established, a distributary channel would continue to be used as any shore parallel barrier spits, which might prevent easy access of fluvial discharge to the basin, will be of diminished thickness and width at these previous sites of fluvial entry when compared with other portions of the coastline. The style of deposition on the delta front would thus have varied laterally with simpler, coarsening-upward sequences developed off

the main distributary channel mouths (section C9, phase 2), and with more complex barred-shoreline sequences between these active sites of fluvial clastic input. The Capel Curig successions thus represent a wave-modified braidplain delta, where both stream effluent and wave action left distinct signatures in the delta-front sand prism (see Fig. 4.54).

If one considers the entire sequence (Fig. 4.37), which generally "shallows" upwards, it is difficult to ascertain whether one should view each shallowing-upward sandstone unit (A-D) as an individual episode of coastline advance, or whether the each unit represents a smaller part of a single prograding shoreline sequence. The thickness of correlative alluvial sequences in the Tryfan anticline, and the upwards decrease in grain size in conjunction with shallowing, favours the former interpretation. In this case, the bottom succession of trough cross-stratified sands could represent nearshore bars. The breaking of storm waves over these bedforms would dissippate some of the basin's deepwater wave energy, and enable deposition of progressively finer-grain sizes as one approaches the coastline (*cf.* Tryfan Fach braidplain delta). Wave ripples would favourably develop in the muddier environments between these shore parallel sand bars; the repetitive lateral migration of the sand bars across interbar areas would create a situation conducive to preservation of the ripples in the geological record.

4.3.4.7 DELTA RETREAT

A thick sequence of delta-retreat facies (phases 8-13, section C9, C3 in Fig. 4.28,4.30) reflects gradual waning of fluvial sediment supply and a concurrently increasing intensity of relative wave energy along the shoreline. The succession fines-upward from conglomeratic coarse sandstone to fine sandstone. The observed facies assemblage, their vertical arrangement and palaeocurrent patterns are all indicative of deposition in a transgressive, high energy, barred shoreface environment (*cf.* Bourgeois,

1980; Leithold and Bourgeois, 1984). Only the lower portion will be described in detail.

Initial evidence for marine conditions (Fig. 4.31) is provided by thin units (section C9, C3, phase 8) of magnetite-rich medium-grained sandstone displaying either trough cross-stratification or horizontal lamination. Units are thought to have been deposited within the upper shoreface of a transgressive wave-dominated barrier island complex. A thin sequence (Fig. 4.30; phase 7) of subjacent siltstone with thin (3-10 cm) tabular beds of fine sandstone, by analogy with a comparable sequence deposited during delta progradation (Phase 5), would represent a narrow lagoon. The preservation of these lagoon facies and the adjacent upper shoreface deposits, and the absence of any evidence for marine erosion (*eg.* gravel or shell lag) during landward migration of the barrier island system indicates that wave reworking and redistribution of shoreline sediments was of limited extent during the marine transgression. A rapid rate of sea level rise (or basin subsidence) in conjunction with a continued sediment supply, is necessary to account for the 'stepwise' retreat of the barrier island(s) (Rampino & Sanders, 1980). The isolated sets of high-angle planar tabular cross-stratification preserved within the overlying ?lower shoreface deposits (section C9) might represent sandridges. In other sequences (Fig. 4.47, base), opposed sets of landward (section C12A) and seaward dipping (section C12B) lamination, probably define large low-amplitude swash bars. Palaeocurrents from trough cross-stratification within this phase of shoreline retreat indicate onshore (NE), longshore (SE or NW) and offshore migration of the lunate megaripples (Fig. 4.31).

These shoreface deposits are overlain by a distinctive conglomeratic horizon (phase 9 in Fig. 4.28, Fig. 4.30; Fig. 4.47) which can be traced for at least 800 metres obliquely transverse to an inferred NW-SE oriented shoreline (Fig. 4.54). The 6-8 m thick unit is thought to represent a

cross-section through a gravelly nearshore bar complex. By analogy with the progradational sequence at Cwm Clorad Isaf (see above), normally-graded, small cobble to granule grade, matrix supported conglomerates (Fig. 4.48), which occur in beds 10-40 cm thick, are interpreted as high density turbidites from alluvial flood or storm-surge ebb flows. They often infill 20-30 cm channels (rip-channels) and increase in abundance, grain-size and thickness towards the inferred shoreline (to the NNE). Conglomerate (clasts up to 10 cm) also occurs as thin horizontal lags which represent redistribution of available gravel on the bar-top by wave processes; imbrication (Fig. 4.48), where present, indicates that the strongest current flow (*ie.* wave swash) is towards the coastline. Trough cross-stratified medium- very coarse-grained sandstones (set size 30-50cm; beds up to 2 m thick) often contain small (1-3 cm) pebbles along the base of foresets (Fig. 4.48). Palaeocurrent analysis indicates most represent either offshore (SE) or alongshore (only to the ENE) migration of bedforms. A large solitary set (1 m) of tabular cross-strat with tangential bases to foresets (section C9, 265m) is interpreted to represent migration of a large sand ridge parallel to the shoreline. Evidence for storm activity increases upwards within the thin conglomeratic horizon, with undulating lamination (C9, 266.5m; C12B, 6m) or more rarely large wave ripples ($\underline{\wedge}$ 40 cm, crestline oriented WNW-ESE) preserved within a thin siltstone horizon (C12A, 3.4m) becoming more prevalent. A progressive decrease in the rate of fluvial input rather than changes in wave regime is invoked to explain the upward increase in the ability of storm processes to modify these nearshore sediments. A planar low angle (10°) scour which cuts at least 1 metre into these wave reworked deposits, contains large scale (1.0-1.5 m) low-moderate angle planar cross-strata (Fig. 4.51). Although outcrop precludes recognition of channel geometry in 3 dimensions, the bedform is thought to represent infill of a narrow distributary channel. An analogy

is drawn with similar channel fills within the Capel Curig Mbr. at Cwm Clorad Isaf (Fig. 4.46). In this case, however, a downcurrent increase in foreset angle from 20° to 40°, in attendant with decreases in maximum grain size from 10 cm cobbles to medium sand, indicates a progressive waning of flow velocity (Jopling, 1965).

The conglomeratic sand bar, is gradationally overlain by a 20 metre thick sequence (Fig. 4.30, 4.31; phase 10) which gradually fines upward from medium sandstone to very fine-grained sandstone. Lithofacies change up-section from undulatory horizontal lamination or swaley-cross stratification (Fig. 4.52) to horizontal planar lamination with low angle discontinuities. The consistency of grain size in the above succession indicates a continual supply of sediment from a largely unchanging source. The sequence is thus interpreted as a transgressive mouth bar. Most sediment was probably directly introduced by fluvial processes, with moderate amounts of wave and storm reworking producing the observed sedimentary structures. Rare tabular to locally channelised horizons of normally graded conglomerate would reflect rip-currents from more severe storm occurrences; this sediment was probably derived from pebbly beaches or nearshore bars on the retreating coastline.

The remaining delta-retreat facies (Fig. 4.28, appendix 5, section C9, 290-360 m; section C3, 80 -130 m) are dominated by a poorly organised succession of intercalated sandstone and mudstone. The thickness and proportion of sandstone beds generally decrease upwards. Most sandstone occurs as thin (3-15 cm), tabular units. Internal stratification within the thicker beds includes normal grading, planar horizontal or low angle lamination with beds often displaying Bouma type (particularly T_{AB}) sequences (*cf.* Brenchley *et al*, 1979). Magnetic heavy minerals frequently dominate horizontal lamination. The lack of evidence for any wave reworking indicates that most of these thin sandstone beds were deposited in

offshore/prodelta areas beneath storm wave base.

The common occurrence of swaley-cross stratification within subjacent mouth bars, however, suggests that storm wave activity was instrumental in their genesis (*cf.* Goldring & Bridges, 1973; Brenchly *et al*, 1979; Tyler & Woodcock, 1987). They might, for example, represent alluvial flood events arising from storm-related rainfall in catchment areas of the drainage basin. A more favourable means of deriving sediment, however, is by liquefaction and retrogressive slumping of mouth bar sediments induced by increases in pore pressure from cyclic loading of the substrate by storm waves (Postma, 1984; Walker, 1985, p.295). Schwab & Lee (1988) have recently described slumps and sediment gravity flows on the Gulf of Alaska continental shelf (1.3° slope) which they attribute to cyclic loading by storm-waves; they illustrate how liquefaction and redistribution of sediment occurs more readily during a long storm than during the limited number of loading cycles generated by an earthquake. The Capel Curig delta front was probably of similar steepness to the Alaskan shelf during the marine transgression; in addition, it would naturally "want" to retreat in tandem with other portions of the coastline, and would thus be even more susceptible to slumping. Wave loading during storms provides a viable mechanism to allow gradual retreat of a shoreline whilst continually supplying sediment to offshore areas. The absence of transgressive gravel lags within nearshore or mouth bars deposits does not therefore signify an absence of wave erosion during shoreface retreat. Several 30-50cm deep channels which cut these horizontally bedded sediments (*eg.* Fig. 4.53) would then represent rip channels from more intense or prolonged storm events.

Although it is obvious that a waning current deposited the above sandstone beds, it is not certain whether the densities of the sediment gravity flows (*ie.* amount of entrained sediment) and depositional slope

were sufficient to generate autosuspension and develop a sustained turbidity current *(sensu stricto)* . Seymour (1986) has presented convincing theoretical evidence to suggest that surges of autosuspended flows may exist within surf zones on high energy beaches. The total absence of features typically found within turbidites such as flute marks, and current-rippling or convolute lamination (T_c), as well as the internal and lateral heterogeneity, indicates little prolonged unidirectional current flow. This favours deposition of the suspensed sediment from storm surge ebb currents, with development of the upper flow regime plane lamination related to wave processes. Similar storm beds have been descibed by Goldring & Bridges (1973) and Aigner (1985); in both cases the dominant Bouma sequence developed is similarly T_{AB}.

The second facies association found within these offshore deposits consists of thick packages (1-4 m) of medium sandstone dominated by trough cross-stratification (20-40 cm sets); they are thought to represent shoals or shallow marine sandbars. A combination of storm, tidal, and longshore currents is probably necessary to explain their genesis (see Tillman & Martisen, 1984). Their thin size points toward deposition in only moderate depths of water (ca. 60 m) on the inner shelf (Stubblefield *et al* , 1984, fig. 8; Tillman, 1985, p.43). The intimate association of these thicker cross-stratified sandstone horizons with the above storm-graded sandstone beds, suggests that the latter were also deposited in shallow depths of water.

It is remarkable that the thickness, grain size changes, and depositional environments of the deltaic-retreat succession are almost a perfect mirror image of the sequence produced by progradation of the braidplain delta. Transgessive sequences of similar thickness from non-barred storm dominated coastlines have been described by Bourgeois (1980) from Eocene sediments of the tectonically active Pacific margin.

Development of such sequences requires rapid rates of subsidence which exceed the high rates of sediment supply. What is more striking in the Welsh example, perhaps, is the regional differences in the thickness of transgressive sequences which varies from 5 m on the west side to 120 on the east side of the Tryfan anticline and 1 metre at Cwm Clorad Isaf over a horizontal distance of *ca.* 5 km (Fig. 4.28). It is certainly necessary to place successions in the Tryfan and Capel Curig structures on separate fault blocks to explain the amount of differential subsidence found between these localities. The NE trending Llyn Cowlyd fault (Fig. 4.28, 4.54) is naturally implicated. The lineament later shows evidence of being tectonically active during deposition of the Racks Mbr.(Chapt. 5.3.3) and Upper Cwm Eigiau Formation (Chapt. 5.4.1), as well as formation of Pb-Zn mineralization at Cae Coch (SH 776 653). Rotational subsidence of the "Tryfan" fault block relative to basement at Capel Curig trapped all sediment which was delivered to the delta front during retreat of the depositional system to the NW of the Llyn Cowlyd fault (Fig. 4.55). It is less certain whether a structural weakness separated depositional sequences on the opposite sides of the Tryfan anticline; only 4° of fault block tilt is required to account for the observed differences in thickness.

4.3.4.8 INFLUENCE OF VOLCANICITY

The full role of volcanicity in the progradation, aggradation, and retreat of the Capel Curig depositional system is difficult to determine. Sandstone petrography, for example, indicates derivation of alluvial fans at Craig yr Ysfa (section F1) on the NE margin of the basin from a mixed tuffaceous sediment/granitic fault block source with minimal volcanic input (Fig. 3.128). The occurrence, however, of thick sequences of redeposited vitric-rich airfall tuffs and feldspathic crystal-rich ashes of rhyolitic composition in lacustrine sequences on the west side of the Tryfan anticline indicates that volcanicity must have influenced some phases of

deposition. A volcanic vent to the west or southwest (*ie*. Idwal fault or its intersection with Llanberis fault system) is implicated to source this material. The quartzose composition of sandstones and occurrence of large (up to 15 cm) rhyolitic pebbles at Cwm Clorad Isaf has already been used to suggest an additional volcanic centre to the west on the Llanberis lineament.

It thus appears that rhyolite volcanicity of varied form (volcanic ash, extrusion of domes) aided advance of the depositional system into the basin, particularly during its period (phase 6) of rapid aggradation on the Tryfan fault block. No doubt tectonic changes (*ie*. uplift of bounding fault blocks relative to the Tryfan depocentre) which preceded this volcanicity, facilitated the intitial progradation of the delta. Although the tectonic relief to the SW of the volcanically-active LLanberis/Glyder Fawr faults was probably of similar magnitude (and perhaps even greater) to the total uplift on the Conwy fault system (Fig. 4.55), the size of the associated alluvial fans appears to have been smaller. They did not influence the overall pattern of sedimentation within the basin. Palaeocurrents in sequences C3 and C9, for example, indicate continued derivation of most sediment from the NE (*ie*. Craig yr Ysfa). The main axis of fluvial systems which supplied the delta front sequences in the Capel Curig anticline, must of 'hugged' the SW margin of the Tryfan fault block. Such assymmetric relationships are common within half-grabens (Leeder & Gawthorpe, 1983) where opposite sides of the basin are subsiding at different rates.

The thick transgressive sequence might record a gradual waning in the rhyolite volcanicity along the margins of the basin. The swing in orientation of the coastline to NW-SE during delta retreat, however, indicates that the only remaining source of sediment was the non-volcanic alluvial fan(s) to the NE at Craig yr Ysfa. The rapid cessation of rhyolite volcanicity obviously provoked this change in basin

palaeogeography and allowed the gradual abandonment of this epoch in the basin's history.

4.3.5 MOEL HEBOG DEPOSITIONAL SYSTEM: WAVE INFLUENCED FAN DELTAS

At Moel Hebog, coarsening-upward sequences subjacent to the Pitts Head Tuff (Fig. 4.56) represent stacking of successively more proximal deltaic facies in response to both faulting and fan lobe progradation.

4.3.5.1 DELTA FRONT

The grain size and lithofacies of distributary mouth bar sequences vary laterally along the delta front. Axial mouth bar successions (Fig. 4.56, 4.57, sequence L2, phase 2) coarsen upwards to medium-grained sandstones and abruptly overlie the mud-dominated prodelta deposits (only top part of prodelta logged). In addition to decimetre scale bedding they contain low angle cross-stratification and undulating lamination. Several beds (up to 1.5m thick) display contorted lamination and large fluid escape structures which are probably due to gravity driven deformation on the delta front. Within proximal mouth bar sequences, amalgamated sets of hummocky cross-stratification are common (Fig. 4.58) indicating high nearshore storm energy. Mouth bar sequences lateral to the location of fluvial input (sequence L1, phase 2) are of finer grain size. Sedimentary structures similarly include horizontal or low angle lamination which commonly yield alongshore palaeocurrents. The absence of HCS or undulatory lamination, however, suggest less nearshore wave energy. Greater amounts of bioturbation indicate slower or more episodic rates of deposition.

4.3.5.2 ALLUVIAL PLAIN SEDIMENTS

Delta plain sediments (Fig 4.57, sequence L2, phase 4) are dominated by trough cross-stratified, medium to very coarse, pebbly sandstones (Fig. 4.59) which are interpreted to have been deposited in relatively small, unstable, braided streams. Channel-fills are crudely organised into 3-5m thick, fining-upwards sequences. Planar cross-stratification is rare,

probably owing to constant shifting of channels (pools) and the accompanying destruction of channel bars (Campbell, 1976). The common occurrence of debris-flow beds in the overlying and adjacent, more proximal alluvial deposits (*eg.* Fig. 4.56, sequence L2, phase 6; sequence L1, phase 5) suggest that the fluvial system operated in the middle to distal segemnt of an alluvial fan.

The fan fringe or periphery was the most complex depositional environment. A simple coarsening upwards sequence of very fine to medium sandstone (Fig. 4.56, section L2, phase 3) is interpreted to represent a moderate wave energy shoreface which formed part of a spit barrier. Characteristic sedimentary structures include planar lamination with low angle discordances and occasional sets of trough and planar crossbedding with offshore directed palaeocurrents.

An anomalous heterolithic 25m thick fine grained interval (Fig. 4.57, section L2, phase 5) is interpreted as representing a tidally-influenced lagoon between fan salients which was protected from strong wave influence by such sand spits or swash bars (*cf.* Vos & Eriksson, 1977; Vos, 1981). A basal 4 metre thick unit of horizontally laminated fine sandstone with siltstone rip-up clasts (section L2, 67-70 m) may represent storm washover sheets. The remainder of the succession comprises interbedding of mudstone and siltstone with coarse sandstone. Thin (2-5cm) units of airfall tuff are commonly preserved within mudstone horizons. Sedimentary structures within siltstone or very fine sandstone (82.2 m) include flaser and lenticular bedding with bimodal palaeocurrents and small wave ripples. The thinnest sandstone interbeds (0.5-5 cm) are usually massive and can be traced laterally for distances of 15-30 m (Fig. 4.60). Their geometry favours sedimentation from sheetflows. These could arise from inactive portions of a distal alluvial fan or overbank flooding from adjacent fluvial channels. Thicker sandstone beds contain planar horizontal lamination (67-70 m), low

angle crossbedding (80.4 m), graded bedding (83-85 m), or climbing ripple cross-lamination; they are similarly interpreted as alluvial sheetfloods or crevasse splays. The thickest sandstone beds (50 to 200 cm) contain moderate scale trough cross-stratification (Fig. 4.61) with foresets occasionally being overturned. They obviously represent infill of small ?low sinuosity streams; further field research is needed to determine whether they represent small fluvial, crevasse, or even tidal channels. They increase in abundance upwards and are gradationally overlain by distal alluvial fan sediments (see below) which lends some support to the first interpretation. The above facies assemblage is somewhat similar to the lacustrine/ floodbasin and transitional fan fringe described from the Hornelen basin (Steel & Aasheim, 1978).

In adjacent sequences (*eg.* section L1, Fig. 4.56), the lagoonal airfall tuffs can still be recognized; they occur, however, as angular rip-up clasts within pebbly trough cross-bedding of poorly channelised braided streams. It appears that tectonic changes associated with the volcanicity induced sideways migration of the locus of sediment input into the basin. A local reconstruction of palaeogeography at this time period (Phase 5) using additional sections (squares on map) indicates coalescing coastal fan-lobes at least 2 or 3 kilometres wide. With the fan-lobe radius assumed to have similar dimension, the alluvial system reflects uplift on the Cwm Pennant or related faults to the west. Facies sequences indicate coastal fan progradation into a basin with high nearshore wave energy. As sediment dispersal patterns suggest rather moderate reworking by waves, the lowermost Moel Hebog succession is categorized as a fluvially constructed, wave-modified fan delta.

4.3.5.3 ALLUVIAL FAN SUCCESSIONS

The lagoon sediments (section L2) are gradationally overlain by a thick succession of medium- very coarse pebbly sandstones interpreted to

represent progradation of an alluvial fan (Fig. 4.57). Distal fan sediments (phase 6) comprise thinly (10-50 cm) bedded sheets of fine-medium coarse sandstones which can be traced laterally over the width of outcrop (10's of metres). Sedimentary structures include unstratified, normal grading, horizontal lamination and small sets of trough cross-stratification (5-10 cm sets). The considerable variation in lithofacies may be viewed in terms of changes in the sediment/water ratio (*cf.* Tryfan Fach alluvial fan at Craig yr Ysfa, Chapt. 4.3.2.4). The development of small channels with tractional streamflow (St) would reflect periods of reduced sediment supply; other lithofacies would reflect deposition by unconfined sediment gravity flows.

Mid-fan sediments (phase 7), in contrast, display less variation in either facies or sediment calibre. Trough cross-stratification dominates sedimentary structures. It is organised into moderate-scale (1-3 m) cycles which display upward decreases in both grain (medium to fine grained) and set (25 to 10 cm) size. Although small (1-2 cm) rounded pebbles of rhyolite do ocurr, sandstone petrography (*eg.* Fig. 3.58, 3.111) indicates sediment was largely derived from basalt and/or mudstone terrains. The amount of basaltic detritus incorporated within sediment appears to increase upwards; sandstones immediately subjacent to the Piits Head Tuff are dominated by Fe-rich oxides (haematite or limonite) and silicates (chlorite) (Fig. 3.63). In adjacent exposures (*eg.* section L1, Fig. 4.70), the coarser grain sizes and larger size of trough cross-stratification (up to 90 cm) indicates development of deeper channels (Cant& Walker, 1978), and suggests a closer proximity to the fan axis. A reconstruction of palaeogeography during maximum progradation of the alluvial fan (*ie.* top of Moel Hebog Mbr.) is shown later (Fig. 4.71). Sequences within the Cwmystradllyn and Y Braich Mbrs. on Moel Hebog indicate a significant change in basin tectonics (Fig. 4.71d) preceeding eruption of the Pitts Head Tuff and will be

discussed in reference to the ash flow sheet iteslf (Chapt. 6.2.5).

4.3.6 MOEL DDU DEPOSITIONAL SYSTEMS: A BIT OF EVERYTHING

Depositional successions at Moel Ddu can be divided into two stratigraphic units (Fig. 4.62, 4.63) which contrast in the direction and composition of their source terrains. The sedimentology of the Lower Moel Ddu Mbr. will be described here and related to correlative sequences at Moel Hebog and Yr Arddu; environmental interpretations of the Upper Moel Ddu Mbr. will be introduced later in reference to the Pitts Head Tuff (Chapt. 6.2.5). Lithofacies assemblages within the Lower Moel Ddu Mbr. can be examined in terms of 3 distinct depositional episodes (A to C) interpreted to represent either coarse-grained deltaic (A,C) or alluvial plain (B) environments. Emphasis will be placed on the middle episode (B) where rhyolitic volcanicity produced continual variation in the grain size and amount of sediment available to fluvial systems. A brief discussion of the lower portion will first be provided, however, to establish the depositional context of the later alluvial sedimentation.

4.3.6.1 A COARSE-GRAINED BRAID DELTA

This first episode of sedimentation (Fig. 4.62, 6.63) coarsens-upwards from siltstone to granule-conglomerate, and is thought to represent progradation of coarse-grained delta. Its second phase (2) of deposition displays little variation in grain size; it consists of horizontally bedded fine-medium sandstone which is thought to have been deposited on a distributary mouth bar. The absence of any wave influence (in comparison with the first phase) indicates a more persistent discharge of sediment into the basin. The channelised units of coarse pebbly sandstone(phase 3) displaying either trough (30-40 cm sets) or planar cross-stratification (70 cm set; M3, 69.7m) would reflect deposition within distributary channels on the lower delta plain. The fine-coarse sandstone which overlies these deltaic deposits (M3, 76-78.5m) indicate several directions of bedform

(sand dune: St) migration; they are thought to represent upper shoreface deposits from a barrier island complex, and indicate a significant change in basin palaeogeography. The abrupt abandonment of this delta lobe was directly associated with a period of rhyolitic volcanicity. The large volumes of fine ash and siltstone introduced onto the lower delta plain initially resulted in the aggradation of a thick series of upper flow regime sheetfloods (Sh). The first indication in the geological record of a forthcoming transformation in the style of sedimentation is provided by a thin (30 cm) inconspicuous horizon of slumped siliceous siltstones deposited within an associated ?interdistributary bay environment (M3b, 74.34 m). Large scale (2-3 m) slump and water escape structures (Fig. 4.63b) represent yet another manifestation of this event, and verify that substantial tectonic (*ie.* fault movement) activity was associated with the rhyolitic volcanism. The deformation signals the culmination of the first outpourings of volcanic ash onto the lower delta plain and "broke the ground" for onset of a new depositional regime.

4.3.6.2 A VOLCANICALLY INFLUENCED ALLUVIAL PLAIN

The middle portion of the Lower Moel Ddu Mbr. is a manifestation of a subtle interplay between volcanicity, tectonics and sedimentation on a "quivering" alluvial plain. Four contrasting facies associations can be recognised which correspond to braided stream, alluvial sheetflood, meandering stream, and lacustrine depositional settings. The existence of any particular environment depended on the nature and amount of associated rhyolite volcanicity.

Very coarse-grained conglomeratic sandstones (phase 5) display poorly developed trough cross-stratification and are poorly organised into 2-3 metre fining-upward cycles. Small (6 cm) sub-rounded to sub-angular rhyolite pebbles and larger (up to 20 cm) rip-up clasts of siltstone often demark foresets. The variation in grain size and texture within sequences

suggest deposition within a hierarchy of small braided streams (*cf.* Tryfan Fach Mbr., Tryfan anticline, Chapt. 4.3.2.3). They also favour derivation from epiclastic weathering of tuffaceous sediment, or access of fluvial systems to several different pyroclastic source regions, rather than direct association with a single episode of explosive rhyolitic volcanicity. In the Garth tuff of the CCVF, for example, sediment only varies in grade from medium sand to coarse silt within the entire 28 km^3 deposit. The coarse grained braided streams within the Lower Moel Ddu Mbr. would thus represent a prolonged period of volcanic quiescence.

During sustained episodes of explosive rhyolitic volcanicity, voluminous quantities of silt or very fine-grained sand were contributed to the alluvial plain. The arrival of this increased sediment load eventually resulted in a profound change in the pattern of fluvial sedimentation. This transformation is well illustrated by sediments (Phase 6A) overlying the previously described braided stream deposits (Fig. 4.64). Thin (10-15 cm) debris flows of conglomeratic sandstone intercalated with floodbasin siltstones foretells the change in basin dynamics and onset of rhyolite volcanicity. This volcanic activity soon expressed itself (Fig. 4.64; phase 6A, unit 2) on the delta plain by thin (ca. 5-10 cm) hyperconcentrated flood flows of siliceous siltstone which are interbedded with units of medium sandstone of similar thickness and origin. The overlying 1.5 m thick unit (3) of very coarse-medium grained sandstone is thought to represent infill of a low sinuosity braided channel; deposits bear some resemblance to sequences from the previous braided streams. The persistent arrival of fine volcanic ash on the braidplain, however, eventually overwhelmed the pre-existing drainage system with all sediment being transported (Fig. 4.64; Phase 6B) as dilute sediment gravity flows or sheetfloods. The transition in fluvial style spans a mere 4 metres within the depositional record (phase 6A, units 5-8; Fig. 4.65).

The thick (5 m) succession of thinly bedded very fine (5 cm) sandstones (phase 6B) which overly these transitional sediments is thought to have been deposited entirely deposited by sheetfloods (Fig. 4.66). Sand units are either massive or contain planar and undulating horizontal lamination, or more rarely display climbing ripple cross-lamination. Lithofacies would represent rapid suspensions sedimentation, upper flow regime plane bed, and migration and aggradation of small current ripples respectively. The lack of variation in grainsize prevents identification of either tops or bottoms to larger depositional units and recognition of any cyclicity (*cf.* Steel and Aasheim, 1978, p.16). The units could either represent sheetflows from distal areas of an alluvial fans or catastrophic overbank flows related to a major trunk stream. With the small size of any channels, it is hard to visualize these thick "background" sequences of very fine sandstone resulting entirely from overbank flooding. Evidence for the contemporaneous alluvial fans needed for the first interpretation is provided within coarsening upwards sequences on Moel Hebog, 6 kilometres (present geography) to the northwest (Chapt. 4.3.5, see below). The deposition of these muddy sheetfloods was directly associated with a period of concurrent volcanicity as indicated by numerous thin (1-5cm) airfall tuff beds. The abundance of dewatering structures attest to the rapid rates of sedimentation.

Several stacked sandstone bodies (Fig. 4.63; phase 6C) intercalated with these sheetfloods (phases 6B, 6D) are interpreted as infill of mixed-load streams of moderate sinuosity. An illustrative example records yet another change in the style of fluvial sedimentation. A thin (20 cm) subjacent debris flow contains abundant angular clasts of rhyolitic air fall tuff (Fig. 4.67) and demarks the end of sheetflood deposition. It can be traced laterally for 20 m before it is cut out by the scoured base of the overlying "meandering" channel. The lower part of the channel fill

consists of cosets of trough-cross stratification with minor undulating lamination. (fig. 4.69a). A distinct package of lateral accretion, with 10-15cm bedding units inclined at a 5° angle, occurs within the top part of the sandstone body (Fig. 4.69b). Accretion surfaces (Fig. 4.69c) are marked by mud or silt drapes. The most common internal sedimentary structure observed is climbing-ripple cross-lamination which yields palaeocurrents normal to the direction of lateral accretion. The channel fill is capped by a laterally continuous 40 cm thick bed of horizontally laminated fine sandstone, which in turn is overlain by a thin unit of floodplain siltstones. The observed sedimentary structures strongly suggest that the sand body resulted from both vertical aggradation and point-bar deposition within small high-sinuosity streams from a muddy source terrain. The basal trough-cross bedding would likely reflect migration of sand dunes within deeper portions of the fluvial channel (*cf.* Jackson, 1976). The alternation of mudstone and sandstone bundles on the accretion units suggests large scale fluctuations in discharge or sediment supply (*cf.* Puigdefabregas & Van Vliet, 1978). The capping unit of horizontal lamination is interpreted as the product of upper flow regime plane bed transport within high velocity flows. The upward increase in flow regime probably reflects shallowing of the fluvial channel which would eventually result in lateral expansion of any flood events into "overbank" areas to form thin, low-depth sheetflows.

The development of these sinuous channel patterns signifies an hiatus in the production of volcanic ash, when discharge was able to erode deep "trenches" into the underlying deposits (*cf.* Vessell & Davies, 1981, their inter-eruptive phase). This episode of stream incision is merely an attempt by the fluvial system to reestablish its former grade. The cohesiveness of the fine-grained ash, by limiting both the bank erodibility and the availability of a local sediment supply, will favour development of

meandering rather than braided channels (Carson, 1984). Similar point-bar deposits of meandering rivers are incorporated within sheetfloods in distal fan associations described by Tunbridge (1981) and Van Meulen (1983).

Lacustrine successions (Fig. 4.63, phases 4,7,9) are the predominant facies association on the alluvial plain. They consist entirely of very fine sandstone, and display either horizontal bedding or fine horizontal lamination (Fig. 4.68). Bedding units are of comparable thickness to those displayed by sheetflood deposits (Phase 6A,6B) and a similar origin is envisaged. Subaerial sheetfloods would flow unimpeded in the low wave energy lacustrine environment (hyperconcentrated flow) and continue travelling as high density turbidity currents further into the basin (*cf.* Sturm & Matter, 1978). Bioturbation is the only clue for a subaqueous rather than subaerial environment. At outcrop sandstones appear less siliceous than their subaerial counterparts. This is thought to reflect a slight reduction in the rate of volcanicity during their formation. The slow "sagging" of the basin which would accompany such a reduction in the rate of magma supply might account for the development of the lacustrine environments.

4.3.6.3 A COARSE-GRAINED DELTA and BASIN PALAEOGEOGRAPHY

The top portion of the Moel Ddu Mbr. (Fig. 4.62, 4.63; phases 9-11) reflects advance of an alluvial plain into nearshore (section M3) and shallow ?lacustrine (section M2) depositional environments. The corresponding wave energy within the receiving basins increased to the north, as reflected by the size of wave ripples and abundance of undulating lamination within "deltaic" sequences. Alluvial plain successions (phase 11A, 11B) include both infill of both high (meandering) and low (braided) sinuosity channels with the proportion of the latter channel types increasing upwards.

Palaeocurrents from fluvial channel-fill sequences within the Lower

Moel Ddu Mbr. (Fig. 4.62) indicate a gradual change in the direction of progradation of the depositional systems from the SE (phase 3), through east (phase 5-8), to NE (phase 10,11). Correlation of sequences between Moel Hebog, Moel ddu, and Yr Arddu (Fig. 4.70) enables reconstruction of local palaeogeography during the three varying episodes of deposition (A,B,C) of the Lower Moel Ddu Mbr. as depicted in Figure 4.71 (a to c).

Sequences within the Upper Moel Ddu Mbr. (Fig. 4.70) and the contemporaneous Nantmor Mbr. (Fig. 4.72) record an abrupt change in basin palaeogeography (Fig. 4.71d) preceeding eruption of the Pitts Head Tuff (see also Chapt. 6.2.5). Facies associations indicate deposition in small coarse-grained, ?fan-deltas which prograded into low-wave energy lacustrine environments. Palaeocurrents from fluvial deposits (Fig. 4.70, 4.72) indicate uplift to the E or SE, with tectonic activity on the Nantmor and Yr Arddu faults probably producing the topography necessary to source alluvial fans. A possible cause for this local reversal in the direction of basin fill will be introduced later (Chapt. 6.2.5)

4.3.7 CAE'R GORS and GWASTADNANT DEPOSITIONAL SYSTEMS: WAVE DOMINATED FAN DELTAS

The Cae'r gors and Gwastadnant Mbrs. of the Lower Cwm Eigiau Fm. (Fig. 4.73, 4.74) are characterised by thin alluvial sequences (10m) of conglomerate and conglomeratic sandstones which overly poorly defined deltaic sequences (50-70m) of fine to medium sandstone. Six successions from two independent depositional systems, at Cae'r gors and at Gwastadnant will be described for illustration. The deposits occur directly below Lower Pitts Head Tuff outflow sheet.

4.3.7.1 DELTA FRONT SUCCESSIONS

At Cae'r gors (Fig. 4.73), the delta front succession comprises two coarsening-upward sequences, both representing progradation of high wave-energy shoreline dominated by shoaling-wave processes (*cf.* Clifton *et al*, 1971; Hunter *et al*, 1979; Clifton, 1981). The first sequence (Phase 1)

coarsens up to medium sandstone, dominated by 1-2 m thick sets of horizontal or low angle lamination with ubiquitous magnetic heavy minerals. Absence of trough cross-stratification may be a function of the finer grain size. The observed stratification types are attributed to high-power (relative to grain size) oscillatory currents, and are though to represent bedforms superimposed on low-amplitude swash-formed bars. Similar deposits have been described from the margins of both modern and ancient fan deltas (McGowen, 1971; Vos & Eriksson, 1977). The second delta front sequence (Fig. 4.73, phase 2) represents new, more pronounced, progradation of the system. It coarsens upwards to pebbly, medium/coarse sandstones with some associated conglomerates. Offshore (SE) palaeocurrents from trough cross-stratification probably represent redistribution of delta front sediments by rip currents.

At Gwastadnant (Fig. 4.74, section E6), the composite delta-front represents similar, repetitive progradation of a shoreface (delta front) environment. The deposits are largely siltstones. Sedimentary structures are largely obscured; remnants of thin (10-20 cm) bedding, rare horizontal lamination, and ripples are the only recognisable features. Periodical influx of coarser sediment is indicated by thin graded beds of very coarse-grained sandstone. The sequence is thought to represent distal to medial mouth-bar deposits, heavily affected by bioturbation. They are unconformably overlain by a thin, strongly erosive alluvial sequence (Fig.4.74, 4.77). The absence of 30-40 metres of expected intermediate-proximal depth mouth deposits is attributed to redeposition associated with tectonic uplift to the NW of the Idwal fault preceding eruption of the Pitts Head Tuff (see below).

4.3.7.2 ALLUVIAL LITHOFACIES

The alluvial deposits in both fan deltas (Fig. 4.73, 4.74) indicate deposition by sediment gravity flows in largely unchannelised distributary

systems. Conglomerate lithofacies, however, vary between the two alluvial fans:

1)At Cae'r gors, matrix supported conglomerates dominate (Fig. 4.75) These are either ungraded or inversely (coarse-tail) graded with gravel fraction comprising less than 10% of deposits. Maximum bed thickness is 100 cm. Clasts are moderately well rounded, and have a maximum clast size of 10 cm. Clasts lithologies (greater than 1cm size, n=80) include, in decreasing order of abundance, fine-grained acidic tuff or tuffite (50%), medium sandstone (27%), basalt (5%), siltstone (3%), and quartzose sandstone (2%). The amount of basalt and siltstone forms a significantly higher proportion of detritus within finer grain fractions.

2)In contrast, clast supported conglomerates dominate the alluvial fan at Gwastadnant (Fig. 4.77). Beds are inversely graded to ungraded, often with matrix rich bases, and occasionally display a(p)a(i) clast imbrication. Clasts are commonly elevated above the tops of bedding units. Largest observed clast size was 33cm. Clast lithologies are dominated by rhyolite and acid tuffite. Bed thicknesses range from 80 to 200cm.

Both conglomerate lithofacies are interpreted as debris flow deposits, and show limited evidence of channeling on outcrop scale (20m). The beds are attributed to laminar sediment flowage with effective shearing restricted to the flow base (ungraded deposits) or a thicker lower zone (inverse grading). Larger clasts probably settled out from the shearing zone due to a combination of dispersive pressure and loss or reduction of strength in the flow matrix. The common lack of grading in the upper portion of beds likely indicates the existence of a non-sheared plug. There is no significant relationship between thickness of beds (Bth) and maximum particle size (MPS). This does not negate, however, a mode of emplacement by mass flows; the lack of correlation would reflect the small size of the data set, fluvial erosion between debris flows and most importantly

amalgamation of flow units.

The textural difference between mass-flows in the two alluvial fans is attributed to their derivation from different source areas (Fig. 4.73, inset). The debris-flow beds are often associated with unstratified, normally-graded or ungraded units (10-15 cm) thick of medium to coarse, pebbly sandstones. These are interpreted to have been deposited by fluidal sediment flows, or hyperconcentrated "sheetfloods", where turbulence and high sediment concentration (high apparent viscosity) were probably the dominant grain-support mechanisms (*cf.* Nemec & Steel, 1984).

In the two fan deltas, the demarcation between subaerial and subaqueous facies is not obvious. A thick sequence of trough cross-stratified, medium-grained sandstones underlying the alluvial deposits in the Cae'r gors sequence (phase 3 in Fig. 4.73) could either represent braided stream or shoreface environments. The evidence of some shoreward (NW) palaeocurents favours the second interpretation. Assuming that alluvial deposits are solely represented by debris-flow or sheetflood facies, the mapped lateral extent of these facies (*ca.* of approximately 800 and 300 metres for the Cae'r gors and Gwastadnant deposits indicate that preserved alluvial fans were extremely small.

4.3.7.3 PATTERNS OF SEDIMENT DISPERSAL

The patterns of sediment dispersal on the delta front are revealed by analysis of stratigraphically equivalent strata in adjacent sequences. Approximately 3 km off the Cae'r gors fan axis, along the depositional strike, the Pitts Head tuff is underlain by a thick (100m) fine-grained coarsening-upward succession (Fig. 4.73, section J2). Its upper, sandy portion is organised into 4 coarsening-upward sequences, each reflecting an episode of shoreline progradation. Sandstones in upper parts of these sequences display parallel, seaward-dipping (ESE) lamination and low angle planar and shallow trough cross-strata, all containing abundant heavy

minerals. These facies indicate a nearshore environment with high wave-energy and considerable periods of sediment reworking by wave swash (*cf.* Reid & Frostick, 1985). This regressive succession could either represent wave-driven longshore drift of material derived from the Cae'r gors fan-delta or be a distal segment of another wave-dominated delta lobe (*cf.* Heward, 1981). Correlative cyclicity (Fig. 4.73), similar heavy mineral contents, andd the short distance involved favour the former interpretation.

At Llyn cwm-y-fynnon (Fig. 4.78), some 3 km basinwards off the Gwastadnant fan, the Pitts Head Tuff is directly underlain by a bipartite coarsening-upwards sandy succession interpreted to represent progradation of storm wave-dominated shoreline. The first portion(A) coarsens upwards from mudstone to coarse-grained sandstone. Offshore deposits consist of very fine-grained sandstone ocurrring either as thin (10-20 cm) normally graded beds or thicker units (20-80cm) containing fine horizontal lamination with low angle disconformities. Sediments of the inferred lower/mid shoreface comprise alternations of undulating horizontal lamination or swaley-cross stratification within fine-medium grained sandstones or trough cross-stratification within very coarse grained lithologies. Palaeocurrents from cross-strata indicate bedform migration towards the inferred coastline, and suggest an origin from shoaling waves for this lithofacies. The overlying succession contains large scale low angle seaward-dipping lamination (Fig. 4.78, 4.79a) which is thought to reflect deposition from storm waves on the upper shoreface or a large coastal sandbar. The preservation of storm beds in this depositional setting implies extremely high rates of sedimentation. Ocassional sets of contained high-angle cross-stratification (Fig. 4.79b) often display complex internal bounding surfaces which cyclically scoop into previously deposited foresets and strata below the set. The 'scalloped' cross-strata

proabably reflect migration of smaller bedforms (*ie.* 'backset' cross-strata) across the lee-slope or along the troughs of the larger shoreward-migrating bedform (Rubin, 1987, Fig. 7). The superimposition of bedforms in this manner is common within wave-dominated nearshore environments where shoaling waves, migration of sand bars both towards and along the shoreline, and offshore relaxation of storm surge currents interact in a complex fashion.

The second shoreline sequence (Fig. 4.78, phase B) displays a similar suite of lithofacies. Several scours filled by fossiliferous cross-stratified medium/coarse grained sandstones are thought to represent rip-channels. The inferred mid-shoreface successions, immediately subjacent to the Pitts Head tuff, comprise alternations of undulating horizontal lamination or swaley cross-stratification within medium sandstone and low angle trough cross-bedding or large straight-crested wave ripples (53cm) within very coarse sandstone (see similar features in terminus of Capel Curig braidplain delta at Cwm Clorad-Isaf; Chapt. 4.3.4.6). The greater amounts of bioturbation and common presence of shelly macrofauna suggest comparatively slow rates of deposition in this second shoreline succession.

The differential response to high-energy shoaling waves within both episodes (A,B) of coastline advance was probably a function of sediment grain size. Available palaeocurrent data suggest that the shorelines prograded due to redistribution of sediments delivered to the Gwastadnant delta front. Both this depositional system and the one at Cae'r gors would then represent wave dominated fan deltas. The higher rates of sedimentation displayed by the first succession (A) can be attributed to uplift and redeposition of the mouth bar deposits which are absent from fan-delta sequences at Gwastadnant (Fig. 4.78). The second episode (B) would reflect the more 'normal' style of delta front sedimentation to be associated with progradation of the Gwastadnant depositional system.

4.4 DISCUSSION

4.4.1 SEDIMENTATION PATTERNS AND CONTROLLING FACTORS

Palaeogeographical configurations varied between deltaic systems, with major changes in sediment dispersal following volcanic eruptions and emplacement of voluminous ash-flow sheets (Orton, 1986; Chapt. 6). The contrasting patterns of delta front sedimentation are a consequence of the varying mode of interaction between sediment-laden alluvial systems and basinal marine or lacustrine processes (Fig. 4.80). Since possible tidal deposits are rare within the studied successions, and where present suggest microtidal conditions (Chapt. 5.4.2), the spectrum of delta types primarily reflects a competition between fluvial and wave processes.

Nearshore wave energy must have been influenced by the size, shape, and depth of the depositional basin. Water depths can be roughly estimated (Fig. 4.80) from the thicknesses of individual deltaic sequences (Klein, 1974), especially where the sediment supply exceeds basin subsidence and where rapid coastal progradation preserves the entire delta-front prism. This appears to be the case here, where alluvial (delta-plain) facies cap all deltaic sequences. Several episodes of deltaic progradation, however, are thought to have been facilitated by concurrent volcanotectonic uplift. Regional tumescence preceding the eruption and emplacement of the voluminous Pitts Head Tuff was particularly important, and the pre-tumsescence water depths of the Cae'r gors and Gwastadnant fan deltas are substantially underestimated (see applied corrections in Fig. 4.80). The amount of error will be related to the magnitude of the volcanic event and sequences from the Tryfan Fach and Magnetite Mbr. braidplain-deltas probably provide more reliable estimates of palaeodepths. These depositional system advanced onto a flat shallow fault block where frictional attenuation reduced deep water wave energy to moderate levels on

the inner shelf. All other deltas appear to have prograded into moderately deep basins, where wave energy was moderately high to extreme depending on coastline orientation. Largest wave ripples in all parts of the stratigraphy (Chapt. 5.5) indicate that maximum wind fetch was from the present south/southeast. The slightly lower nearshore wave-energy confronting the Gwern Gof depositional system can thus be explained by the delta facing obliquely away from this direction of main wave attack.

The inferred local variations in overall wave climate within the basin, however, appear rather insignificant when compared with the actual spectrum of delta types. The contrasting delta morphologies and associated patterns of sedimentation (summary in Fig. 4.82) are thus primarily a consequence of differences in fluvial processes. The critical parameters which controlled these deltas and determined their differing styles of development are thought to have been the geometry and slope of the alluvial "feeder" system, the rate of sediment supply (discharge magnitude and temporal variability), and the grain-size of sediment delivered to the delta front. These physical parameters varied among and between the alluvial fans and braidplains (see below), and dictated the varying responses of the associated deltas to basinal wave-climate (Chapt. 4.4.3)

4.4.2 THE ALLUVIAL COMPONENT

4.4.2.1 CONTROLS ON CHANNEL PATTERN

The complexity of channel styles and deposit types exhibited by alluvial sequences, can be related to the variety of partly independent controls that govern fluvial sedimentation. Recent developments (Ferguson, 1987) have shown that more emphasis must be placed on the amount and nature of sediment supplied as a control on channel pattern, than traditional approaches which would stress hydraulic controls as governed by slope-discharge thresholds. The amount of water discharge can now be simply viewed as determining the dimensions of the channel (width, depth), but the

relative proportions of bedload (usually sand and gravel) to suspended load (usually silt and clay) will control not only the channel shape (width/depth ratio) but channel pattern (Schumm, 1981). If the combined sediment load from upstream and bank erosion, for example, exceeds transport capacity, shoaling will cause braiding; meandering will occur where both bank erosion and upstream sediment supply are low relative to flow strength (Schumm, 1963; Carson, 1984). Braiding, which involves a greater amount and rate of channel modification than meandering, requires a greater stream power and therefore steeper slopes for any given discharge. With more extreme sediment concentrations, the dampening of turbulence within streams due to interpartical collisions (Leeder, 1983) will result in transport of detritus as sediment gravity flows (*ie.* where the sediment moves the fluid).

This new emphasis on the influence of sediment parameters on channel pattern (bank erodibility, sediment supply, bedload capacity) is directly applicable to volcanic terrains where large temporary sediment loads which might literally 'fall from the sky' into fluvial systems, could wreck havoc with the style of fluvial sedimentation. Changes in channel pattern should be more common within volcaniclastic sequences than within fluvial sediments deposited in non-volcanic terrains. An excellent example of this local control of sediment supply on alluvial sedimentation is provided by depositional sequences of alluvial plain successions at Moel Ddu (Chapt. 4.3.6.2; Fig. 4.63, 4.64). Here, introduction of large quantities of fine-grained ash associated with rhyolite volcanicity resulted in transformation of both braided and meandering channels, and episodes of rapid aggradation of sheetfloods on the alluvial plain. A return to either low or high sinuosity channels depended on the availability and grain size of sediment relative to discharge. Braided streams reflected a moderate supply of heterolithic coarse-grained volcanic ash, whereas development of more

sinuous channels resulted when the only supply of sediment was through incision and lateral bank erosion of channels within previously deposited fine-grained cohesive volcanic ash.

Within braided streams themselves, the channel size and depth, as well as temporal variability in discharge were probably the dominant controls on the type of bedform, and resultant sedimentary structure, that developed. Where the full channel dimensions cannot be recognised, a crude idea of the size of a fluvial system (and amount of discharge) is given by both the thickness of fining-upward channel-fill cycles and the mean height of contained cross-stratification (Miall, 1976; Cant and Walker, 1976). Narrow (*ca.* 150-300 m), 'fixed', bedload-dominated channels associated with large fluvial systems (*eg.* Gwern Gof Mbr.) contained abundant linguoid or transverse bars. Channels were filled by rapid vertical accretion which favoured preservation of bar elements (Sp). It is significant that both delta plain distributaries (this study; Bhattacharyya & Lorenz, 1983; Hazeldine, 1983; Okolo, 1983) and anastomosing rivers (Smith, 1983) display similar channel fill sequences. These are both low gradient, laterally confined, sandy, bedload-dominated channels with steady discharge. It appears that in such situations, transverse bars commonly span the entire bankfull river width and retain their identity during any slight fluctuations in discharge; the rapid aggradation associated with extreme rates of sediment supply and limited channel mobility results in stacked sets of planar cross-strata.

In contrast, in the broader (1-3 km) bed-load dominated fluvial systems (*eg.* Capel Curig Mbr. braidplain), more substantial fluctuations in both sediment supply and discharge (partly stemming from volcanicity), resulted in frequent dissection of transverse bars (*cf.* Allen, 1983) and production of a hierarchy of deeper fluvial channels separated by sand flats (Cant & Walker, 1978; Galloway, 1981). The lateral migration of the

longer-active channels accross the braidplain (Campbell, 1976) would lead to the preferential preservation of their contained dune-type bedforms (St), at the expense of the cross-channel bars (Sp), as observed.

4.4.2.2 CONTROLS ON ALLUVIAL FAN MORPHOLOGY

On a larger scale, the supply of sediment relative to discharge argument can be used to understand complete alluvial systems. Modern alluvial fans, for example, can be divided into two end members categories: fluvially dominated and mass flow-dominated (Bull, 1962, 1972). Fluvially dominated fans are of larger size and lower gradient, and generally display a good correlation between their gradient and size of transported sediment (*eg.* Boothroyd, 1972; Schramm, 1981). This relationship is useful in interpreting ancient depositional sequences. Fans produced almost entirely by sediment gravity flows are smaller and have steeper slopes (Bull,1962, 1972)

Fan morphology is also directly related to drainage basin area (Denny, 1967; Hooke, 1967, 1968; Ryder, 1971; Harvey, 1984). With large catchment areas, more regular discharge results in large, fluvially dominated fans. With small, rugged drainage basins, where streams are ephemeral, fluvial processes are less effective in moving sediment, and hence mass-flows become more important. In addition, the steeper gradients in smaller basins render the sediment stored on slopes susceptible to instability and gravity-induced mass movements during heavy rainfalls (Baker, 1977; Wasson, 1977; Harvey, 1984). Sediment movement on associated fans is extremely infrequent and, depending on climatic conditions, may occur only during flood events every few centuries (Wasson, 1977; Kostachuk *et al* , 1986).

In the Devonian Hornelen basin of Norway, small, steep, mass flow-dominated fan deltas developed along the northern margin can be distinguished from large fluvially dominated fan deltas of the southern margin (Gloppen and Steel, 1981). The differences in fan thickness,

morphology and facies are related to drainage-basin areas, with small alluvial catchments along the tectonically more active northern margin (Steel, 1976). In the North Wales basin, in the absence of any known variation in climate, the recognised diversity of alluvial systems (and resultant delta types) is also attributed to the tectonic control on catchment areas.

The Gwern Gof alluvial fan (Fig. 4.8) contained large, low sinuosity bed-load dominated streams carrying fine- to medium-grained sand. Sediment composition and grain size suggest deposition during a tectonically "quiescent", non-volcanic period, when organised fluvial systems were able to derive sediment from a large drainage basin. Interdistributary bays and coastal lagoons were important components of the flat lower-delta plain.

The alluvial fans at Craig yr Ysfa (Fig. 4.20) represent intermediate alluvial fan types and show evidence for deposition by both streams and mass flow processes. The relative abundance and grain size of these deposits types can be used to infer the relative size of the associated alluvial system (Fig. 4.81). In the Tryfan Fach Mbr., for example, the rapid vertical changes between braided stream and debris flow deposits, suggest association with a small alluvial fan. In contrast, the comparative absence of either debris flows or profound vertical variation in grain size or facies within the Capel Curig Mbr. at the same locality, indicates a larger stream-dominated alluvial fan from a more distant source. An estimate of absolute fan sizes is possible (Fig. 4.81) through comparison of grain size and textural parameters of deposits with trends derived from modern alluvial fans in humid settings (Boothroyd & Nummedal, 1978; Schramm, 1981). The vertical organisation of coarsening upward alluvial fan megasequences (Tryfan Fach, Capel Curig, Moel Siabod Mbrs.) at Craig yr Ysfa indicate both successive back-faulting and repeated uplift of the basin margin at this locality. The absence of any capping large-scale

fining-upward sequences gives evidence for rapid abandonment of fan lobes through avulsion or tectonic change rather than a gradual period of scarp retreat and lowering of relief.

The wave-dominated Gwastadnant and Cae'r gors fan-deltas represent the smallest depositional systems studied. These were associated with small alluvial fans dominated by sediment gravity flows. Development of any delta-plain lagoons was precluded by the steep (*ca.* 5°) fan gradients. In a study of modern, coarse-grained, stream dominated fan deltas on a microtidal, low-wave energy coast of Honduras, Schramm (1981) similarly noted the control of fan slope on lower delta-plain environments, with the delta plain consisting of longitudinal bars. In the Welsh examples, the progradational delta-plain sequences comprise "proximal" debris-flow conglomerates overlying sandy sheetflood deposits, suggesting downstream dilution of sediment flows and decrease in their competence. The difference in type of mass flows found on the Cae'r gors and Gwastadnant fans can be attributed to the position of deposits on the alluvial cone, drainage basin characteristics, and source rock lithology. Sediment gravity flows at Cae'r gors, for example, were deposited on the mid- to distal- portion of the alluvial fan (Chapt 4.3.7.2) whereas debris flows from the Gwastadnant alluvial system were probably situated on a proximal position of the alluvial cone, and unconformably overlie distal to intermediate-depth mouth bar deposits. In addition, the increased comparative size of the Cae'r gors alluvial system is thought to reflect discharge from a correspondingly larger drainage basin; the greater abundance of water available to fluvial systems enabled transport of most sediment as hyperconcentrated sheetflows rather than viscous debris flows.

A secondary control on the morphology of all alluvial fans, is the nature of the bedrock lithology at source (*cf.* Bull, 1962). This was particularly important in small depositional systems such as the Cae'r gors

and Gwastadnant examples, where sediment was supplied from a limited number of lithogies. The marginally larger size and finer-grained sediment calibre of the Cae'r gors alluvial system, when compared with its neighbouring fan at Gwastadnant, must therefore be partially a consequence of derivation (Fig. 3.128) from a basalt/mudrock as opposed to quartzose volcaniclastic/ rhyolite pluton source. By the same reasoning, the Gorllwyn Mbr. alluvial fan at Moel Hebog was probably of lower gradient than comparable depositional systems (*eg.* Capel Curig Mbr.) at Craig Yr Ysfa. This lower slope allowed development of broad shallow lagoons on the fan margins between apices of dominant fluvial input.

4.4.2.3 ALLUVIAL FANS VERSUS BRAIDPLAINS

The contrasting depositional styles of alluvial fans and braidplains is best seen by comparison of individual alluvial systems between Craig yr Ysfa and the south side of the Tryfan anticline. The braidplain sequences (Fig. 4.13, 4.14, 4.32) display less vertical (and hence downstream) variation in both grain size and attendant facies (predominantly St) when compared with correlative alluvial fan successions, even though the braidplain was influenced by concurrent rhyolite volcanicity to the W and SW. The comparatively flat braidplain allowed widespread lateral migration of the low sinuosity streams to form thin (2-4 m) tabular sheets (200-500 m) sheets of sandstone. The flat topography also allowed development of numerous shallow lakes between the main fluvial axes. Within the alluvial fan sequences, outcrop prevents proper analysis of the degree of lateral variation in lithofacies. The amount of scouring visible on a local scale, however, suggests that channel incision followed by vertical aggradation was a more prevalent mode of fluvial deposition than found on the braidplain.

The homogeneity of fluvial sequences on braidplains, when set against variation displayed by alluvial fan sequences, suggests that spatial and

temporal variability in both discharge and sediment supply were comparatively insignificant. Braidplains received sediment from several alluvial fans on opposite sides of the depositional basin. As a result, their supply of detritus was not seriously interrupted when either tectonic change or a cessation of local volcanicity led to retreat of individual "feeder" alluvial systems. In addition, since topographic contours on braidplains will tend to parallel the coastline, avulsion is not a viable mechanism for abandoning the entire depositional prism, as it appears to be on many alluvial fan and meandering stream depositional lobes. The derivative braidplains thus all indicate gradual reductions in sediment supply associated with their "abandonment". This expressed itself in the Tryfan Fach and Capel Curig examples by substantial thicknesses of delta-retreat facies which overly braidplain deposits. It is probable that many thick transgressive shoreface to offshore sequences found in the geological record (Levell, 1982), similarly record retreat of braided coastal plains.

4.4.3 DELTAIC SEDIMENTATION

In trying to understand processes of nearshore sedimentation on a delta front, it is useful to discriminate between dissipative and reflective shoreline morphologies (Wright *et al*, 1979; Short and Hesp, 1981; see Chapt. 5.3.5). Dissipative coastlines will have low overall gradients, a complex 3 dimensional topography with common longshore bars cut by rip currents, and waves breaking a considerable distance from the shoreline. In contrast, reflective beaches will be characterised by smooth 2-dimensional nearshore profiles with most wave energy reflected directly from the beach itself. Abundant fine sediment, a wide shallow inshore profile and high wave energy will usually produce dissipative conditions; a coarser grain size, reduced sediment supply and lower wave power will increase reflectiveness at the beach face.

4.4.3.1 FAN-DELTAS VERSUS BRAIDPLAIN DELTAS

The contrasting depositional styles of fan-deltas and braidplain-deltas can be considered and interpreted against the background of the physical factors outlined above. This is best revealed by a comparison of two superficially similar, wave-modified deltaic systems: the Moel Hebog fan-delta and the Capel Curig braidplain-delta (Fig. 4.80). Both systems delivered sediment of similar size grade to the delta front by mobile braided streams, and both prograded into a basin with high nearshore energy. They differed, however, in both their overall geometry, their nearshore evolution, and their ability to cope with marine energy.

The Moel Hebog fan delta (Fig. 4.56; 4.71), was characterised by a lobate or arcuate geometry, partially inherited from its associated alluvial fan, as well as a lower, more 'flashy' sediment supply, as the system relied solely on one drainage basin as a source. The common occurrence of debris-flow deposits suggests that the alluvial plain of this delta was of a steeper gradient; Rust(1972, 1979) has noted the higher slopes (0.019) of braided fluvial systems from the Spring Creek tributary alluvial fan when compared with those (0.006) from the associated Donjek braidplain. To some extent, it is the alluvial system that determines the inshore slope of the delta front, especially the part which is above modal wave base. For example, Dunne and Hempton(1984) examined two contrasting styles of deltaic sedimentation in the pull-apart Lake Hazar basin and noted that an alluvial slope of at least 0.05(3°) was critical to produce fan deltas with steep 0.50 slope(28°) Gilbert-type foresets. Lower gradient braidplains in the axial zone of Lake Hazar basin resulted in fine grained low angle(1°) "mouth bar-type" deltas (see also Wood & Ethridge, 1988).

In the Moel Hebog fan-delta system, dissipative conditions were associated with mouth bars; rip-current deposits resulted from wave-induced coastal set-up (*cf.* Wright, 1977; Wright *et al*, 1980). However, a

combination of smaller overall sediment input and steeper inshore slope on the fan delta resulted in reflective beach faces, characterised by simple 2 dimensional coarsening-upward shoreface profiles along much of the delta front. With waves dispensing energy directly on the shoreline due to steeper gradients and with a general deficit of inshore sediment storage, a reflective shoreline is likely to experience greater erosion on the beachface (Wright *et al*, 1979).

The arcuate geometry of fan deltas will further assist wave processes in the redistribution of delta front deposits. Wave refraction has two effects. By reducing wave height, it enhances the development of reflective beach morphologies, and by reducing the obliquity of wave incidence, it concentrates wave power on the fan salients. The curvature of the delta front will also induce longshore drift regardless of the angle of incidence of dominant wave-attack to overall shoreline orientation. A good modern example of such drift patterns is provided by the Jaba delta in Papua New Guinea where normal wave incidence results in bidirectional sediment movement on either side of an axial river mouth (Wright *et al*, 1980). In the Moel Hebog fan delta, the sediment from such longshore drift is inferred to have been distributed and accumulated laterally between fan salients as barrier spits. Due to this pattern of sediment dispersal, delta progradation was comparatively ineffective; depositional sequences farther to the NE (*eg.* at Cae'r gors) are silt dominated.

In contrast, the Capel Curig braidplain-delta (Fig. 4.54) was probably fed laterally by several tributary fans, one of these being represented at Craig-yr-Ysfa. A combination of higher sediment supply and lower inshore slope would result in the fully dissipative barred coastline, which has been inferred from facies sequences. Although dissipative shorelines have high rates of seaward sediment transport (Wright *et al*, 1979), their large inshore sediment storage creates the greatest potential for net subaerial

accretion. With any (*eg.* seasonal) reduction in incident wave height much sediment will be returned to the shoreface (*eg.* Clifton *et al*, 1971) by migration and welding of nearshore bars and ridges (*eg.* Hine & Boothroyd, 1978). Although such bars will influence the location of fluvial input, as in the Capel Curig example, they will not inhibit progradation of the delta itself as a sheet-like prism.

In addition, Short and Hesp (1981) have noted a genetic relationship between subaerial beach topography, aerodynamic flow regime and potential landward aeolian sediment transport. Such sediment movement was most favoured on dissipative shorelines with resultant formation of large beach foredunes. Facies associations of such aeolian bedforms have not yet been found within braidplain deltaic sequences in North Wales. They were likely an important component, however, of delta plain aggradation as they are on modern glacial outwash plain shorelines (Hine & Boothroyd, 1978). Being backed by active braided streams, however, they would have a low preservation potential in ancient sequences.

Good modern analogues for the contrasting Moel Hebog and Capel Curig deltas are perhaps provided by the Yallahs fan delta of Jamaica and the Skeidararsandur coastal braidplain of Iceland (Fig. 4.84). Both these depositional systems have seasonal discharge in a network of pebbly braided streams and prograde into high wave-energy marine basins and insignificant tidal influence on the pattern of sedimentation. In the Yallahs example, owing to the coarse size of sediment, limited sediment supply, and steep inshore slope, waves expend much of their energy directly against the shoreface. Most of the delta front is thus of erosive nature, with a steep coarse grained foreshore zone. The delta coastline only prograded downdrift of the river mouth where refracted waves tended to parallel bathymetric contours; here, the lower wave energy in combination with continuous supply of longshore derived sediment resulted in broad sandy depositional beaches.

In contrast, a more extreme wave-energy, a much greater sediment supply and a lower inshore slope have caused the beach and nearshore zones of the Skeidararsandur delta to develop a complex rhythmic topography that includes coastline barrier spits and large offshore bars. Dissipative nearshore conditions, combined with an oblique incidence of maximum wave-energy resulted in substantial seaward and longshore sediment transport. In spite of this, a 75km stretch of shoreline, was able to advance 1.1 kilometres in 300 years (Hine & Boothroyd, 1978). Occasional glacial burst floods caused by subglacial volcanic eruptions assisted fluvial processes by periodically causing catastrophic floods with discharges of 50000 m^3/s ('normal' peak summer flow was 469 m^3/s). Progradation of this depositional system could produce patterns of delta-front sedimentation similar to the Capel Curig braidplain delta example, with mouth bars reflecting localised sediment input through distributary channels and shore-parallel bars and shoals characterising the rest of the delta front.

4.4.3.2 FAN-DELTAS: THE IMPORTANCE OF ALLUVIAL REGIME

Fan deltas are a consequence of progradation of alluvial fans into standing bodies of water. The size and morphology of fan deltas are largely dictated by the nature of the contributing alluvial fan. In the Devonian Hornelen basin of Norway, alluvial fans prograded into a shallow, low-energy floodbasin lake (Steel & Aasheim, 1978; Pollard *et al*, 1982), the dominant influence of this passive body of standing water was subaqueous dilution of debris flows and a variety of "gravity-winnowing" processes on water-saturated subaqueous slopes (Larsen & Steel, 1978; Nemec *et al*, 1984).

In North Wales, however, deltas prograded into a basin with high wave-energy. They reveal that the nature of fluvial input, as determined by tectonic controls on catchment area, was crucial in dictating the effectiveness of the delta in coping with marine energy and in controlling

the extent of sediment reworking by waves, and the type of resultant coastline.

In the Gwern Gof fan delta, the high rates of sediment supply in relation to basin energy enabled progradation of large fluvially-dominated deltas. The comparatively fine sediment calibre and inherent low gradient delta slope assisted progradation by dissipating wave energy. The wave dominated fan deltas at Gwastadnant and Cae'r gors represent the other end of the spectrum, being associated with small sediment gravity flow dominated alluvial fans. The extreme variation in grain size delivered to the delta front resulted in a spectrum of coastline morphologies. Low sediment supply, coarser grain size and steeper inshore gradient, however, favoured development of fully reflective shorelines. As a result of infrequent sediment input, which must have characterised these fans, wave processes were able to redistribute most of the sediment delivered to the delta front, with dispersal both offshore and alongshore. The finer grained, relatively passive, distal portions of the mass flow-dominated alluvial fans would be particularly susceptible to marine erosion and redeposition. The preservation of the thin "proximal" alluvial sequences in the Cae'r gors and Gwastadnant fan-delta examples may be, in fact, only a consequence of early burial by emplacement of the overlying Pitts Head Tuff soon after their deposition.

4.4.3.3 BRAIDPLAIN DELTAS

The three braidplain-deltas examined differ in the amount of lateral variation displayed by delta-front sequences. With the Tryfan Fach and Magnetite Mbr. depositional systems, the high rates of sediment supply, combined with effective attenuation of wave energy on a low gradient muddy continental shelf, allowed synchronous progradation of the entire braidplain as a fluvially dominated sheet across the previous shoreline. Although regional distribution of lithofacies suggests that the braidplain

and related alluvial fans of the Capel Curig depositional system were considerably larger than those associated with the Tryfan Fach and Magnetite system, progradation was not directly a consequence of volcanicity, with most sediment transported reflected reworking of previously deposited volcaniclastics due to tectonic uplift. The more sporadic supply of sediment, assisted by a higher wave-energy on fault blocks to the SE of the Llyn Cowlyd lineament (Chapt. 5.5.2), allowed significant wave reworking of fluvially-derived detritus. Related formation of barrier spits forced braidplain discharge through several discrete distributary channels resulting in progradation of a more varied coastline with both shore parallel sand bars and mouth bars.

4.4.3.4 TOWARDS A CLASSIFICATION

When interpreting ancient coarse-grained deltas, one should first attempt, if possible, to discriminate between fan-deltas, braidplain-deltas and braid-deltas (as defined here). The examples from North Wales reveal the importance of the alluvial system in determining: (1) the effectiveness of discharge in coping with marine energy; (2) the pattern of delta-front sedimentation; and (3) the development of either reflective or dissipative shorelines. Several authors (*eg.* Rust & Koster, 1984, p. 164; McPherson *et al*, 1987) have also discussed the contrasting tectonic implications of these various types of alluvial systems.

In fan deltas, the size grade and abundance of sediment will be a critical factor in governing coastal response to wave energy and resultant patterns of delta-front sedimentation. Sediment grain-size will also strongly influence the slope of the delta front. It is not surprising that most examples of small fan deltas, especially those on which sediment gravity flows are the dominant alluvial process, have been described from lacustrine settings or from extremely small basins where subaqueous slopes are not significantly reworked by wave processes (*eg.* Larsen & Steel, 1978;

Postma, 1983; Link, 1984; Nemec *et al*, 1984). Even where protected by patch reefs, the low rate of sediment supply and coarse sediment calibre allows development of erosive reflective shorelines (Hayward, 1985; Dabrio & Polo, 1988). The inherent steep slopes in such fan deltas will also promote subaqueous redistribution of delta-front deposits by slumping and gravity mass flow (*eg.* Postma, 1983, 1984). Deltaic cyclicity will often result from progradation of the entire alluvial fan, rather than from deposition related to seaward decrease in discharge momentum and grain size of sediment transported by channel effluent (as in "ordinary" deltas). At the other end of the spectrum, large alluvial fans, owing to their finer-grained and more abundant supply of sediment, will be more efficient in coping with basin energy. Large scale growth faulting will be a dominant deformation feature on the delta front.

The morphology, sediment distribution and facies patterns of both fan and braidplain deltas is controlled by competing fluvial, tidal, and wave processes. It is reasonable, therefore, that deposits should be classified according to the dominant process (Kleinspehn *et al*, 1984) as in the case of "ordinary", river deltas (Galloway, 1975). However, to better understand coarse-grained deltas, and perhaps all delta types (Reading, 1986, p. 521), the ternary classification diagram of common deltas should be extended in a fourth dimension to account for the grain size of sediment delivered to the delta front along the axis of the alluvio-deltaic prism (Fig. 4.83). A similar approach has recently been used to comprehend shallow marine siliciclastic facies (Johnson & Baldwin, 1986, p.258). This makes it possible to compare deltas will similar gradients on the alluvial plain and delta front, similar responses to nearshore wave energy, and similar catchment areas and rates of clastic supply. The various types of alluvial system will overlap along the "sediment grade" axis, or the fourth dimension, of the extended classification diagram (Fig. 4.83).

Chapt 5. SHALLOW MARINE ENVIRONMNENTS

> " The world is not divided into sheep and goats. Not all things are black nor all things white. It is a fundament of taxonomy that nature rarely deals with discrete categories. Only the human mind invents categories and tries to force facts into seperated pigeonholes. The living world is a continuum in each and every one of its aspects. "
>
> *Kinsey, Pomeroy, and Martin,*
> *Sexual Behaviour in the Human Male, 1948*

5.1 INTRODUCTION

Shallow marine environments, both shelf (offshore) and nearshore (shoreface) settings, are influenced by a complex set of physical processes (Fig. 5.1), in particular fairweather and storm processes (*eg.* Hobday & Reading, 1974). Fairweather processes are "the normal day to day processes operating on the sea floor and include not only deposition of mud from suspension, bioturbation and wave activity, but powerful tidal and oceanic currents [whereas] storm processes are catastrophic events which include wave-, wind- and gravity-driven currents" (Elliott, 1986b). In addition, the response of marine environments to these processes (*ie.* facies or sub-environments that develop) is strongly dependant on the sand-mud content of available sediment, and rate of sediment supply (Fig. 5.2).

Shallow marine facies, with no obvious deltaic associations, form the bulk of Caradocian sedimentary successions in North Wales. Although this has long been recognised, there are few detailed published descriptions of sedimentary facies, notable exceptions being observations from the Berwyn Hills (Brenchley & Pickerill, 1980) and South Shropshire (Hurst, 1979; Brenchley & Newall, 1982). The contained shelly fauna have historically attracted most interest, with researchers erecting palaeoecological community models (Fig. 5.3) (*eg.* Pickerill & Brenchley, 1979; Lockley, 1980; see review in Lockley, 1984). The success of these faunal "communities" or associations, however, in identifying and discriminating between depositional environments has been far from spectacular. It now

appears (Janusson, 1984; Lockley, 1984) that the texture (grain size) of substrate exerted a strong control on faunal distribution (Fig. 5.4) in the Ordovician. In non-volcanic settings (*eg.* Ashgill or Silurian in Wales), where 'normal' hydraulic processes are free to operate and Walther's law is applicable (see Chapt. 8.2), the grain size of the substrate will largely correspond to the energy of the environment (*ie.* available food) and broad shore-parallel zones of benthic communities (*eg.* Ziegler, 1965) may be recognised. This contrasts with basins influenced by active volcanicity where the texture of the substrate often bears no strict relation to the energy of the environment, but is merely a manifestation of the sediment grade produced by local capricous outpourings of volcanic ash. Stratigraphic changes in grain size will be largely a manifestation of variation in the source parameters (*ie.* type and amount of volcanicity) rather than different depositional environments or water depths. It is thus not surprising that no clear pattern has emerged from mid-Ordovician palaeoecological studies. Intergradation between faunal associations combined with problems in identification of brachiopod species due to tectonic deformation, resulted in most faunal assemblages (samples) collected from the study area being placed in a ubiquitous "Dalmanellid-sowerbyellid **megacommunity**" (Wright, 1979), which appears to span most depositional environments.

As a result of these complexities, this analysis of shoreface and shelf sequences will place little emphasis on available knowledge regarding faunal associations (Wright, 1979; BGS internal reports, Appendix 2), but instead concentrate on interpretation of sedimentary structures (facies) and depositional sequences (facies associations). After a brief review of some processes which appear to have operated within Caradocian shallow marine environments, broad interpretations of several depositional sequences will be presented to document the mode and pattern of sand

dispersal within this portion of the Welsh basin. Facies will later be used to provide some constraints on bathymetric parameters (water depth, basin size) and comment on the environment of deposition of the intercalated mud component.

5.2 IDENTIFICATION OF SHALLOW MARINE PROCESSES

There are few criteria which are unequivocably diagnostic of particular "fairweather" or "storm" marine process. Trough cross-stratification is a notable example, and could arise from either shoaling nearshore waves (Clifton *et al*, 1971), longshore currents, storm surge ebb currents, tidal currents (Johnson, 1977), storm-related geostrophic flows (Tillman & Martison, 1984), or even traction currents within high density turbidity currents (Lowe, 1982; Hein, 1987). "Herring-bone" type cross-stratification, which was once thought to be diagnostic of tidal currents, has now been described from both upper shoreface deposits where it is attributed to lunate megaripples from shoaling waves (Roep *et al*, 1979) and from a tributary junction of modern anastomosing fluvial channels (Alam *et al*, 1985).

The origin of hummocky cross-stratification (Fig. 5.5) is equally enigmatic. The structure was originally attributed solely to vigorous oscillatory flow generated by storm waves (Harms *et al*, 1975), with bedforms assumed to display NO systematic direction of migration. Most researchers now believe, however, that a combination of waning storm-generated unidirectional flows with superimposed oscillatory storm wave action is necessary for formation of these bedforms. Arguments have been based either on analysis of ancient sequences (Hunter & Clifton, 1982; Dott & Bourgeois, 1983), on observations from modern shallow marine bedforms and hydrodynamics (Swift, 1983; Swift & Nummedal, 1987), or theoretical considerations (Allen, 1985). Convincing field evidence for a combined-origin for HCS-type bedforms, has recently been provided by Nottvedt and

Kreisa (1987) who describe low angle cross-stratification from the Cretaceous Carolinefjellet Fm.(Spitsbergen, Norway) which displays 'classical' HCS-type structures when viewed transverse to inferred flow directions, but a preferential dip of foreset laminae in longitudinal sections (parallel to flow). They recognise (Fig. 5.6) a series of bedforms related to a complete spectrum of oscillitory to unidirectional currents. Symmetric high-energy oscillatory currents will yield plane parallel stratification as one end member (high suspended sediment load transport) whereas purely unidirectional flows produce angle-of-repose cross-stratification (bed load transport). The low angle cross-stratification and 'classical' hummocky cross-stratification represent the middle portion of this spectrum. The former will, of course, be more common within small fault-bound, tectonically active basins (*eg.* this study, Eocene Coledo Fm., Chan & Dott, 1986; Lower Jurassic Hasle Fm., Denmark, Surlyk & Noe-Nygaard, 1986) where gravity-driven density (high sediment load) currents will often assist storm waves in redistributing liquefied nearshore sediment (Chapt. 4.3.4.7). In topographically flat epicontinental seas (*eg.* Cretaceous interior seaway, references in Duke, 1985) such gravity assistance will be less common. Coarser grain sizes will also favour development of low-angle rather than hummocky megaripples in response to combined storm flows.

Deposits from both types of combined flow megaripples, termed LACS (low angle cross-stratification) and HCS (hummocky cross-stratification), are observed within shallow marine successions in North Wales. In one case, the two bedform types were observed in the same depositional unit (Fig. 5.7A, 5.7B) supporting speculations by Nottvedt and Kreisa (1987, p.360) that only a slight reduction in the relative strength or directional consistency of the unidirectional component is necessary to result in formation of hummocky rather than low-angle megaripples. A detailed analysis of one example of each type of combined-flow megaripple is

presented below, to illustrate the care which must be taken when interpreting "undulatory" lamination within shallow marine sequences.

A hummocky cross-stratified storm deposit within nearshore sediments of the Capel Curig braidplain delta (Fig. 4.46) overlies a sharp planar to slightly undulating contact. Although some individual laminae do thicken slightly into the crests (see Brenchley, 1985, Fig.7; Hunter & Clifton, 1982, Fig. 6B), most depositional units are bound by subtle low angle truncations and actually represent drapes of laminae over an irregular hummock and swale topography (see Harms *et al*, 1975; Bourgeois, 1980). The close association of both accretional and erosional elements within its genesis supports suggestions that the area of existence for hummocky cross-stratification on flow regime diagrams occurs along the boundary between sheetflow (sand movement) and 3-d ripple fields (P.Allen, 1985, Fig.1; Dott & Bourgeois, 1982, Fig. 20). Regardless of the exact mechanism of bedform growth, the upwards transition from near horizontal laminae to increasingly convex forms indicates an upwards reduction in the oscillatory flow stress and transition from upper flat bed to lower flow regime conditions (Harms *et al*, 1982). The top surface of the hummocky bedform is well preserved with a relief of 20 cm, with local modication by small wave and current ripples developing discontinuous pebble lags and small-scale trough cross-stratification.

An excellent example of a low-angle megaripple (Fig. 5.8) is provided by a 20-70 cm thick storm bed from the Upper Cwm Eigiau Fm. at Bwlch Trimarchog, which can be traced for over 100 metres parallel to the inferred flow direction. The base of the storm unit periodically scours 20-40 cm into subjacent strata (Fig. 5.8b) with a trough-to-trough spacing (*ie.* bedform wavelength) of 3-6 metres. Laminae are characteristically horizontal or display low-angle dips of less than 15° (Fig. 5.8b) towards the inferred downcurrent direction (north). Internal truncation surfaces

are particularly common and are often overlain by higher-angle foresets more typical of trough cross-stratification(Fig. 5.8b, c). The top of the storm bed displays low-angle megaripples (Fig. 5.8b) with wavelengths of 1 to 1.5 metres, which in turn contain small wave ripples on their stoss and lee sides. The geometry and internal structures of the above sandstone unit are interpreted to result from low relief bedforms that migrated principally in one direction under the influence of combined wave surge and unidirectional currents as a result of a single waning storm event.

Low angle undulatory cross-strata has also been described, however, from graded beds deposited by stratified turbidity currents (Prave, pers. comm., 1986 cited to Duke, 1987) on a submarine fan in 500 metres of water. The similarity between the combined flow low-angle megaripples and subaerial pyroclastic "surge" deposits is perhaps even more striking (compare Fig. 5.9, 5.10 with Fig. 5.6 - 5.8). Both bedforms involved 1) imposition of progressive gravity waves (internal body waves within stratified flows for pyroclastic "surges"; Valentine, 1987) onto undirectional currents and 2) a combination of suspended-load and bed-load sediment transport (see Sigurdsson *et al*, 1987 for pyroclastics). Several examples of undulatory low-angle cross-stratification within coarse-grained volcanic ash found within ??subaqueous deposits of the 4th volcanic mbr of the Capel Curig Volcanic Formation (Fig. 5.11; see also Chapt. 7.2.4) or Bedded Pyroclastic Formation (Fig. 5.12) are attributed to the influence of internal waves developed within surging subaqueous sediment gravity flows rather than water-surface waves related to ?concurrent storm activity. Undulatory or hummocky cross-stratification must therefore always be examined within its depositional context, before any inferences regarding its genesis can be made.

Additional problems arise in interpretations drawn from storm-generated hummocky cross-stratification, when one considers the size of the

bedforms involved. 'Classical' HCS appears to grade downward in size to morphologically similar structures with smaller spacings. Hummocky bedforms with spacings of 0.30 to 0.60 cm have been described, for example, from lacustrine upper shoreface (water depth less than 2 m) environments (Greenwood & Sherman, 1986) where they are attributed to a combination of primary wave oscillation and a secondary 'steady' longshore current; even smaller microhummocky lenses (Dott & Bourgeois, 1982) or truncated wave ripple laminae (Campbell, 1966; de Raaf *et al*, 1977) have long been noted from offshore facies. In Wales, small scale "hummocky cross-stratification " can be seen within 10-15 cm thick sandstone beds within the late Ashgill Scrach Formation near Llandovery in the Towy Anticline, or within fine offshore sandstones of the Moel Siabod Mbr. at Rolwyd (Fig. 6.6b, SH 7656 5086). The definitional 1 metre lower limit assigned by Harms *et al* (1975) to hummocky cross-stratification is thus entirely arbitrary (see Duke, 1987). Although all the above bedforms are genetically related, the size of the waves involved and water depths that one would infer would vary greatly. The small scale bedforms, for example, could be formed by modest waves in shallow environments (*eg.* Greenwood & Sherman, 1986), or the influence of larger waves from intense winter storms or sub-tropical hurricanes in deep water (*ie.* close to storm wave base). A large part of the controversy that has arisen over identifying the type of storms responsible for 'classical' hummocky cross-stratification (Marsaglia & Klein, 1983, Klein & Marsaglia, 1987; Duke, 1985, 1987) must surely arise from failure to consider the spectrum of bedform sizes that are involved.

In summary, although it now appears to be certain that a combination of unidirectional flows and wave oscillation is necessary to generate hummocky cross-stratification, in subaqueous volcanically-influenced terrains one must be extremely careful to acsertain whether the gravity waves involved represent internal waves within a stratified subaqueous

sediment gravity flow or water surface waves related to storm activity. Within the latter type, it is ludicrous to infer either minimum water depths (as done by Walker, 1984) or type of storm involved (*cf.* Marsaglia & Klein, 1983; Klein & Marsaglia, 1985; Duke, 1985), solely from the presence of hummocky cross-stratification. The sedimentary structure thus only: A) provides a record of the existence of wave-driven currents (*ie.* storms of varying magnitude) in dispersing sediment within a basin and, B) serves as an indicator of "storm" wave base within depositional sequences.

As a consequence of the above complexity, full comprehension of shallow marine successions is only attainable through analysis of facies associations rather than individual sedimentary structures. Patterns will more complex in proximity to shorelines where the modification of wave and tidal circulation patterns creates confusion. These nearshore environments will be considered first.

5.3 SHOREFACE SEQUENCES

5.3.1 INTRODUCTION

Two contrasting transgressive and one progradational non-deltaic shoreface depositional systems will now be examined to illustrate the influence of (1) physical regime (wave & tidal processes), (2) sediment supply, and (3) tectonic setting (bathymetric changes) on shoreline morphology. These sequences will be later compared with deltaic shorelines (Chapt. 4) to establish a general model for shoreline sedimentation within the study area.

5.3.2 LOWER RACKS MBR.: A Transgressive barred coastline

The Lower Racks Mbr. (Fig. 5.14) arose from reworking of the loose unlithified top portion (1/3) of the 40 m thick Garth tuff, during the marine transgression (Chapt 6.2.2) that followed its emplacement; shard textures (*eg.* Fig. 3.18) are visible in all thin sections examined from this stratigraphic unit. The reworked deposits often lie with sharp

disconformity upon the basal welded portion of the Garth tuff (Fig. 5.14), which suggests that the top of the ash flow sheet formed a low relief coastal plain which was erosively "planed-off" before deposition of the Lower Racks Mbr.

The first sedimentary structures observed within the Lower Racks Mbr. (Fig. 5.13, sections B3, A1, A2, A3) include low-moderate angle cross-stratification (Fig. 5.16), vertically-aggrading wave ripples (Fig. 3.14), undulatory horizontal-to-low-angle lamination, and horizontal bedding. Successions are thought to have been deposited within a wave-dominated shoreface. Any faunal assemblages are dominated by *Macrocoelia exspansa* (Wright, 1979, p.143) which reflects both the high wave energy and fine substrate (Fig. 5.3, 5.4; Pickerill & Brenchley, 1979). An isolated set of desiccation cracks and small ?current ripples (Fig. 5.18, section A2, phase 2, 6.25 m, Fig. 5.13) which overlie wave reworked deposits indicates that the subjacent shoreface deposits formed part of a barrier island complex (Fig. 5.22). One does not have to infer progradation of a coastline to account for development of this coastal feature. The sudden introduction of large quantities (ca. 2 km^3 spread as a 10-20 m thick sheet over 100 km^2) of fine-grained unlithified volcanic ash, in a area previously noted for its high wave energy (Chapt. 4.3.4.6), would naturally lead to the upwards growth of sand bars **without** any additional sediment supply from land. It is not certain, however, whether these offshore bars could actually emerge from wave-produced aggradation to produce the barrier island complex. Most researchers (*eg.* Swift, 1975 cited by Reinson, 1984) consider that submergence of beach ridges is a more likely scenario. This genesis is not inconceivable in this case, however, as the Garth tuff **must** have formed a subaerial coastal plain (Chapt 6.2.2) for a significant time after its emplacement before it subsided into the stormy Welsh seas. It is interesting to note that Davis (1986) has recently suggested that barrier

islands preferentially coincide with distinct breaks in topographic slope, and one must wonder whether the Llyn Cowlyd fault (Fig. 5.22) was influential in this regard.

The overlying, 10 m thick succession (Fig. 5.13, section A2, phases 3,5) of very fine-grained thinly bedded sandstone (Fig. 5.19), with little variation in grain size between depositional units, is thought to represent amalgamated storm washover sand sheets. Evidence for the inferred lagoon setting (and hence the barrier island setting; Fig. 5.22) is provided by a thin siltstone unit (Fig. 5.20, section A2, 20 m) which contains both dalmanellid and sowerbyellid brachiopods in addition to the ubiquitous ramose bryozoa (Fig. 5.15; Wright, 1979, p.139). The predominance of storm washover sheets within lagoon-fill sucessions suggests not only that the barrier island complex was of low topographic relief, but that it was cut by few tidal inlets (*ie.* microtidal conditions) to accommodate landward-directed waves and wind-forced currents during storms.

The next phase of coastal deposition (Fig. 5.13, section A2, A3, phase 6) is interpreted as a transgressive shoreface. Trough cross-stratified medium-coarse grained and horizontally laminated fine-grained sandstone characterise the upper shoreface whereas horizontally bedded, bioturbated, fossiliferous fine-medium grained sandstones typify deposits from the lower shoreface. The only record of the bar top and foreshore environments from the transgressive barrier island which is present in the depositional record is a thin lag (section A2, 27.2 m) of siltstone rip-up clasts which overly a irregular (10 cm of relief) erosive contact.

A reconstruction of palaeogeography during deposition (phase 3,4) of the Lower Racks Mbr. is depicted in Fig. 5.22. A broad subaqueous mud flat is envisaged to the NW of the Capel Curig anticline, where a thin 2-5 m thick mudstone horizon forms the entire sedimentary interval between the Garth and Racks (2nd volcanic mbr. of CCVF) ash flow tuffs. A subaerial

alluvial "plain" would have existed to the NE of Craig yr Ysfa, where no epiclastic sediment was deposited between the first two volcanic members of the CCVF. Faunal assemblages (BGS internal report PD/80/12) from the "mud flat" deposits are dominated by *Dinorthis* brachiopods on Tryfan, whereas on Gallt-yr-ogof, *Dalmanella* and *Howellites* genera are most common. At the south end of Gallt-yr-Ogof (BGS locality PD/80/12, no.17), the abundance of *Lyrodesma*, semi-infaunal cyrtodontids (bivalves), and *Lingulasma*, which are often in life positions, indicates a stable substrate, low rates of sedimentation, and little post-mortem transport. Although it is not certain that all, if any, of the faunal assemblages belong to the Lower Racks Mbr. (they could be the Middle Racks Mbr.), the increase in environmental energy to the NW within the lagoon could be attributed to reformation of the wind-driven waves behind the barrier island(s). The development of both progradational (Phase 1,2) and transgressive (Phase 3,4) barrier island sequences during a slow marine transgression, would be merely a vagary of differences in storm energy received by the various segments of the depositional system (*cf.* Heron *et al*, 1984).

5.3.3 MIDDLE AND UPPER RACKS MEMBERS: A REGRESSIVE WAVE-DOMINATED SHOREFACE

In the Capel Curig anticline 80 m of epiclastic sediment separate the 1st and 2nd volcanic members of the CCVF (Fig. 2.10b, 5.13, 5.14). The top portion of this sedimentary package comprises the Middle and Upper Racks Mbrs., which are thought to represent progradation of a wave and storm influenced shoreface to coastal mud-flat. Within the Middle Racks Mbr., the distribution of sandstone, sedimentary facies, and palaeocurrents indicate that the source of coarse-grained sediment lay to the W or SW, probabably reflecting uplift along the postulated extension of the LLanberis Pass fault. Sandstone petrography (Chapt 3.4.2, Fig. 3.111) suggests derivation from tuffaceous sedimentary rocks.

The lower shoreface-offshore (Fig. 5.13, sections A1, A4) consists

predominantly of fine- to medium-grained, horizontally-bedded sandstones. Depositional units often display normal grading (Fig. 5.24) in the more distal (to the east, section A3) successions. The lateral variation in unit thickness, lack of current-structures, and close proximity to definitive shoreface deposits argues for a storm origin for these graded sandstone beds (*cf.* Brenchley, 1985).

Further evidence for storm and wave activity is provided within mid-shoreface successions (Fig. 5.13, sections B3, A1 to A4), where 'bell-shaped' scours (Fig. 5.24) thought to represent rip-channels, are filled with shelly coquinas. Trough cross-stratification within scour fill deposits yields palaeocurrents to the south. The scours are often associated with 10-30 cm thick tabular beds which can be traced for distances of 30 m along depositional strike. Fauna in both types of coquina are dominated by *Macrocoelia* and *Salopia* (Wright, 1979), which display no preferred orientation of disarticulated valves. Although one can not be certain that the scour events and deposition of shell beds are genetically related (*cf.* Goldring & Aigner, 1982), storm relaxation flows are still thought to have been important in accounting for shell emplacement. The scattered ocurrences of brachiopods found within adjacent strata, however, suggests that the abundant accumulations of bioclastic debris within shell beds arose through amalgamation of numerous storm events by *in situ* fairweather wave and tidal-current winnowing of the nearshore substrate during periods of reduced sediment supply (*cf.* Bloos, 1982; Fürsich, 1982). This would also explain the absence of plane to hummocky laminated fine-medium grained sandstone (*cf.* Kreisa & Bambach, 1982) which one would expect associated with storm beds.

Isolated sets of planar-tabular cross-strata (Fig. 5.25), with foresets dipping at angles of 10 to 35°, and set sizes up to 1 metre, are also attributed to "fairweather" processes in the mid-shoreface setting.

The size of cross-strata, and consistency of foreset angle within individual sets, argues against a "catastrophic event" and bedforms are thought to represent small straight-crested sand-waves. The exact nature of currents responsible for bedform migration is uncertain, although a weak bimodality impli some tidal influence and ebb-tidal currents might be involved. Additional field research (especially palaeocurrents) is needed to fully constrain patterns of sand dispersal during this time period.

A reconstruction of palaeogeography within the uppermost portion of the Middle Racks Mbr. (Fig. 5.27) illustrates that areas to the NW of the Capel Curig anticline are again thought to have comprised a broad low relief "mud flat", bounded to the SE by the Llyn Cowlyd fault. This fracture, in fact, formed a "hinge-line" which not only allowed greater subsidence in the Moel Siabod fault block to the SE (Howells & Leveridge, 1980), but provided a "shelf-slope break" between the mud flats on the Tryfan fault block and the shoreface sequences in the Capel Curig anticline. To properly understand patterns of sediment dispersal within these nearshore sequences, it is necessary to segregate palaeocurrents derived from trough cross-stratification associated with coquina beds and those from planar tabular cross-strata as shown. It is significant that relaxation currents (St) within rip-channels from coastal set up of storm waves flowed to the SSE (190°); wave ripples from all parts of the stratigraphy (*eg.* Chapt. 4.3.4.6, see also Chapt. 5.5.1) show that the largest storm waves in the Welsh basin during the Lower Palaeozoic came from this direction. In addition to the brachiopods noted above, the mid-shoreface successions frequently contain disarticulated crinoidal columnals (Wright, 1979, p. 144). It is envisaged that crinoidal bioherms or banks were an important component of the mud flat environment, with shelly fauna stripped from the crinoid ramp slopes by offshore directed bottom return flows providing the bioclastic debris found in sequences examined from the

Capel Curig anticline (cf. Aigner, 1985, p. 117-120). A similar situation existed within the Muschelkalk sea of Germany (Hagdorn, 1985) where crinoids inhabited the shallow marginal parts of the basin with brachiopods related to small bioherms in deeper environments. A multitude of tabular winnowed shell lags (Fig. 5.13, section A4, 30-35 m; section B3, 69 m) capping the Middle Racks Mbr. record a waning in sand supply and a subtle change in the patterns of sediment dispersal.

The Upper Racks Mbr. is comparatively unfossiliferous in comparison with other epiclastic successions within the CCVF and consists either of parallel-laminated to pervasively bioturbated very fine-grained sandstone, or more conmmonly structureless mudstone, with the thickness of the mudstone component greatest in the SW. It could be argued that the return to mud deposition reflects basin-wide subsidence prior to eruption and emplacement of the Racks Tuff. It must be remembered, however, that the previous source of sand supply lay to the WSW, whereas the Racks tuff was erupted from a vent to the NE (Fig. 5.28). A more attractive interpretation, therefore, would be that rotational tilt of fault blocks away from the Racks tuff vent interupted patterns of sand supply with the mud component derived from uplifted 'lagoon' successions to the north. Mud is thought to have been deposited in coastal plain to offshore settings.

5.3.4 UPPER CWM EIGIAU FM., PONT-GROMLECH -WAVE-INFLUENCED TRANSGRESSIVE NON-BARRED SHORELINES

At Pont-y-Gromlech, the Pitts Head Tuff (PHT) is overlain by a fining-upward sequence of sandstone (Fig. 5.29, 5.34), interpreted to represent a transgressive non-barred wave-influenced shoreline. Medium- to coarse-grained sandstones with high-angle trough cross-stratification and finer lithologies displaying upper flow regime plane bed lamination (Fig. 5.29, section E3, Fig. 5.30; Fig.5.44, 3-5m; Fig. 5.46) are thought to represent upper shoreface deposits. Lithofacies would be attributed to migration of lunate megaripples from shoaling storm waves, wave swash, and interaction

between oscillatory flow and storm relaxation currents respectively. These upper shoreface deposits infill, steep-sided scours (Fig. 5.30). The "gullies" were probably originally formed by subaerial erosion of fluvial channels into the fine-grained unwelded top portion of the PHT, rather than solely due to rip-currents.

More distal sandstones within the first transgressive sequence at Pont-y-Gromlech (Fig. 5.29, section E3, 15-30m; section E1, 40-65m; section E5, 10-50m) are dominated by parallel-laminated to bioturbated very fine-grained sandstones with minor wave rippling. This facies assemblage is typical of the lower shoreface-offshore transition zone (water depths 9-18 m off California, Howard & Reineck, 1981) or inner shelf (Bourgeois, 1980). Tabular to locally scoured coquina beds, similar to those from the Middle Racks Mbr (Chapt. 5.3.3), are also an important component of the depositional succession. In one notable occurence (Fig. 5.32), the coquina beds form an integral part of a large sand bar complex, and in forming a hardground/bioherm on the sea floor may even have been instrumental in initiating bar development. Winnowed shell concentrations are particularly common on the stoss-side of this bedform (see further discussion in Fig. 5.32). In analogy with the Racks Mbr., the above shell accumulations probably also indicate periods of reduced sediment supply and pronounced wave and/or enhanced tidal current winnowing. The amount of bioturbation present within adjacent sandstones similarly suggests slow rates of deposition. The length of geological time that might be incorporated within the formation of the coquina beds is brought home when one attempts to correlate these shoreface sequences with shallow marine successions at Cwm Cneifion (Fig. 5.33), a mere 1000 m to the NE. At this location, the Upper Cwm Eigiau Formation is 250 metres thick and comprises at least 4 regressive and transgressive sequences. It becomes clear that the shoreface sequences at Pont-y-Gromlech are stratigraphically condensed with

shell beds only representing some of the erosional disconformities present within the succession. A reconstruction of palaeogeography in the vicinity of Llanberis pass (Fig. 5.34), using additional sections from Gwastadnant, Llyn Clyd and Ogwen Cottage, reveals that one must involve both the LLanberis pass and Llyn Idwal faults to account for patterns of basin subsidence after emplacement of the PHT.

5.3.5 CONTROLS ON SHORELINE SEDIMENTATION

The three non-deltaic shoreface successions reveal, as did patterns of delta front sedimentation (Chapt.4) that fairweather and storm wave processes were the dominant factors responsible for redistribution of nearshore sediments. However, whereas the fan- and braidplain-deltas commonly displayed evidence for storm processes (*eg.* wave ripples, HCS) within nearshore deposits, the non-deltaic shorelines, regardless of orientation, were dominated by fairweather wave and longshore processes. It is not neccessary to invoke changes in wave regime within the basin to explain this discrepancy, as the preservation of discrete storm events within shoreface sequences, in a basin which also had high fairweather wave energy, will be dependent on high rates of sedimentation. This condition was not fulfilled in non-deltaic shoreline systems where everyday processes erased most storm evidence.

As mentioned earlier (Chapt. 4.4.1), when trying to understand such wave-influenced coastlines it is useful to discriminate between dissipative (barred) and reflective (non-barred) shoreline morphologies (Fig. 5.35, 5.36). These represent end-members of a complete spectrum of coastline types (Fig. 5.37). Most modern coastlines (Wright *et al*, 1979; Clifton *et al*, 1971) display both temporal (*eg.* with seasons) and spatial (*eg.* Fig. 8 in Wright *et al*, 1979) variation in the type of inshore depositional system. It will thus not be surprising in analysis of the morphodynamics of ancient nearshore systems to record a similar variability. By

attempting to view shoreline morphologies in terms of the above dissipative and reflective types (Fig. 5.38) , however, better insight will be gained into long-term patterns of inshore sediment circulation. With deltaic coastlines (Fig. 5.39), (A) a fully dissipative coastline will be dominated by a **mouth bar** sand prism (*eg.* Mississippi) where most deep-water wave energy is attenuated before the nearshore; the morphologic types of Wright *et al*, 1979 will occupy the middle portion of the complete coastline spectrum, and (B) there will be more lateral variation in nearshore morphology (*eg.* Yallahs fan-delta, Fig. 4.84) than displayed by non-deltaic coastlines due to greater alongshore variation in the rate of sediment supply.

Analysis of wave ripples from shallow marine successions (Chapt. 5.5.1) indicates that that the maximum wave period (and hence wave energy flux) varied from 3 to 8.3 seconds depending on basin configuration or fault block morphology. These different wave conditions obviously had an effect on the type of coastline that developed (Fig. 5.39; Chapt. 4.4.1). This is perhaps best illustrated within the Capel Curig Mbr.. On the west side of the Tryfan anticline, braided streams prograded into low wave energy lacustrine/marine environments and simple coarsening-upward non-barred coastlines (Fig, 4.29, Phase 2A, 2B) developed (*ie.* reflective coastline). In contrast, further to the southeast (east side of Tryfan anticline, Fig. 4.28, 4.30; Capel Curig anticline, Fig. 4.37), the more significant supply of sediment input from additional sources along Llanberis pass and higher wave power resulting in progradation of partially dissipative coastline (Chapt. 4.4.3.1)

Most coastlines prograded, into basins with an unchanging moderately to extremely high (in the Welsh context) deep water wave energy. Differences between and among the deltaic and non-deltaic coastlines can be related to the amount and grain size of sediment supplied. Most of the

non-deltaic shorelines (*eg.* Upper Cwm Eigiau Formation at Pont-y-Gromlech, Fig. 5.29; Middle Racks Mbr., Fig. 5.13) were reflective and dominated by shoaling waves, occasional storm relaxation flows and shore-parallel longshore currents. The Lower Racks Mbr. represents a special case; although obviously not related to any deltaic depositional system, a superfluous supply of fine-grained sediment was provided by the unwelded top to the Garth tuff. A dissipative, barred, coastline was a natural consequence.

Differences between and among the fan- and braidplain deltaic coastlines have already been discussed (Chapt. 4.4.3). What will be emphasised here, is than even **within** a single depositional system one can relate the various coastal responses to wave energy to the grain-size and abundance of sediment supply. During the first episode of progradation (A in Fig. 4.78) of the Gwastadnant fan-delta, for example, the high sediment supply (Fig. 5.39) from redeposition of the proximal mouth bars deposits from Gwastadnant itself resulted in formation of large sand bar complexes on the delta front (Chapt. 4.3.7.3). In contrast, a more simple, coarsening-upward non-barred shoreline dominated by storm wave processes characterised the later (B in Fig. 4.78) progradational phases when 'normal' sedimentary processes operated. Similarly, during **progradation** of the Capel Curig braidplain delta the higher rate of sediment supply resulted in formation of large (at least 40 m thick) mouth bars (Fig. 4.30, phase 2) on the delta front. This sub-environment was a minor component (Fig. 4.30, 4.31, phase 10, only 15 m thick) of the delta front during retreat of the depositional system, when the more reduced sediment supply relative to wave energy permitted more extensive development of nearshore sand bars (phases 9, 11, ?13).

5.4 SHELF DEPOSITIONAL SYSTEMS

Shelf sandstones may be classified (Fig. 5.40) on the basis of their

position on the shelf (shoreface-attached, inner shelf, middle shelf, outer shelf), and on the basis of whether they are deposited during a transgression, regression or stillstand (Tillman, 1985, p. 1). The shallow marine "fairweather" and "storm" processes (Fig. 5.41) which are active will of course vary in terms of location on the shelf. Most tidal currents, for example, will only affect the sea bottom on the inner shelf (less than 30 m water depth).

In the Welsh basin, storm-influenced sandstone lobes have recently been singled out as the dominant shallow marine process in the Ordovician epicontinental sea from both inner shelf and nearshore (Brenchley & Newall, 1982; Brenchley & Pickerill, 1980; Hurst, 1976), and mid-shelf (Tyler & Woodcock, 1987) depositional environments. In this portion of the thesis, storm-depositional lobes from the Upper Cwm Eigiau Fm. will be first examined with the aim of illustrating both the shallow nature of the "offshore" environments, and the volcanic control on bathymetry within basinal successions. Evidence for tides (Chapt. 5.4.3) is then illustrated from two transgressive nearshore sequences.

5.4.1 STORM DEPOSITIONAL SYSTEMS: THE UPPER CWM EIGIAU FM.

The Upper Cwm Eigiau Fm. (UCEF), comprising sediment between the Pitts Head Tuff and the Snowdon Volcanic Group, varies in thickness from zero (0) m to the SW of Cwm Llan to 1500 metres between Bwlch Craig and Pen Llithrig-yr-Wrach (Fig. 5.42), 12 km to ENE, where three depositional packages (X,Y,Z) can be recognised. These profound changes in thickness are related to differing degrees of fault-controlled basin subsidence, with the Llanberis Pass, Idwal-Moel Hebog, and Nantmor-Llyn Cowlyd framework of fault systems being particularly important. In addition to allowing different rates of subsidence between adjacent fault blocks, the tectonic framework also controlled both depositional slope and the positioning of various depositional environments. The absence of any sediment from Cwm

Llan to Moel Hebog, for example, suggests that the Beddgelert fault block was largely subaerial during UCEF time (Orton, 1987). The shoreface sequences from Llanberis pass have already been described; the object here is to examine more offshore successions further to the east (Fig. 5.43).

5.4.2.1 UPPER CWM EIGIAU FM.- Cwm Cneifion/Llyn Bochlwyd.

At Llyn Bochlwyd and Cwm Cneifion, the UCEF consists of a series of poorly defined coarsening and fining upward packages of sandstone and siltstone of 25-50 m thickness (Fig. 5.44, 5.45). They are divided by several "tuffite" horizons which are now thought to represent subaqueous pyroclastic flows (Chapt 3.2.2, 7.2.6). The common presence of almagamated sets of hummocky-cross stratification and undulating lamination (*eg.* Fig. 5.46, section D1, units 2, 5-7; Fig. 5.45, section D2, units 5-7) indicates deposition above storm wave base. Sandstone packages are thought to represent progradation and retreat of storm-dominated sandstone lobes on the inner shelf (*cf.* Brenchley & Newall, 1982). Both regressive and transgressive sequences can be recognised.

The lowermost succession of sandstone and mudstones overlying reworked facies of the Pitts Head Tuff (PHT) is regressive. At Cwm Cneifion (Fig. 5.44), the muddy base of the sequence (unit 2, enlarged section) is abruptly overlain by thin (5-10 cm) beds of horizontally laminated to bioturbate very fine-grained sandstone (unit 2b) thought to represent "distal" storm events. This lithofacies quickly passes upwards into a comparatively thick succession of medium-grained sandstone (unit 2c) containing amalgamated sets of both hummocky and low-angle cross-stratification (Fig. 5.7A, 5.7B). Individual beds (*ie.* depositional units) are up to 100 cm thick with lensoid coquinas of brachiopod valves commonly occurring at the base of shallow scours. The abandonment of the small sand lobe, and attendant reduction in the rate of sediment input, is recorded by a thin capping horizon of bioturbated very fine-grained sandstone, and

tabular, non-graded coquina beds attributed to wave winnowing processes. Similar deposits are found with the condensed shoreface sequences at Pont-y-Gromlech further to the SW (Fig. 5.29, 5.32). Coquinas (dominated by *Dinorthis*) are also found within thin, sharply-based, normally graded sheets within the overlying offshore mudstones where they would represent storm distribution of these nearshore bioclastic concentrates.

The above storm-dominated sandstone lobe can also be identified at Llyn Bochlwyd (Fig. 5.45), 1000 metres to the NNE. Here it consists solely of thin (5-10 cm), isolated, horizontally-laminated to bioturbated or massive

very fine-grained sandstone beds intercalated with mudstone (Fig. 5.47) attributed to the influence of storms below effective wave base. It is obvious from the total thickness of the sandstone package, and in particular the rate of vertical and lateral change in facies (*cf.* Aigner, 1985, Fig. 22; Brenchley, 1985, p.380 where bathymetry and proximal to distal facies changes span 50-100 km), that: 1) sediment supply was limited, 2) the shoreface attached sandstone lobe was <u>not</u> associated with a river mouth, which would produce larger 30-70 metres sequences, 3) the basin deepened comparatively rapidly accross the Llanberis Pass and Glyder Fawr faults after emplacement of the PHT from fairweather wave base (15 m) at Pont-y-Gromlech to greater than storm wave base (ca. 50 m; Chapt. 5.5.1) at Llyn Bochlwyd. If the same bathymetric slope was maintained across the Tryfan fault block, one would infer water depth of ca. 90 m for correlative mudstone facies at Pen-yr-Helgi Du (SH 700 630; Fig. 5.43, composite section F, ca. 380 m). As it is unlikely that changes on the fault block were comparable with those along its tectonically active margins, this value can be taken as an upper limit.

The transgressive successions contrast in all aspects with the regressive sequence desribed above. Firstly, they were deposited during

active periods of rhyolite volcanicity, and intercalated siltstone horizons are all siliceous. Embayed quartz crystals and euhedral feldspar laths are commonly observed; pumiceous tuff fragments are also found. Although storm processes were still the dominant mode of sediment transport, as indicated by thick amalgamated sequences of "scour-and-fill" type structures and hummocky cross-stratification (Fig. 5.45, section D1, unit 6, 305-315 m; Fig. 5.44, section D2, unit 7, 172-180 m; Fig. 5.48), more proximal sandstones also contain channelised (less than 20 m wide and 1 m deep), normally-graded, beds of very coarse to fine-grained sandstone. Beds are either structureless or contain low-angle cross-stratification within their top portions. They represent sandy high-density turbidity currents with deposition from both rapid sediment fallout from suspension (massive sandstone) and late-stage traction sedimentation within residual low-density (finer-grained) currents (*cf.* Lowe, 1982, Fig. 8). This lithofacies decreases in both abundance and grain size upwards as the proportion of storm-generated beds and chlorite content within sandstones increases. This progressive reduction in the rate of sediment input suggests a waning of rhyolitic volcanicity.

All transgressive sequences overly primary pyroclastics. A model can easily be envisaged for the transgressive primary pyroclastic/volcanic ash units (5,6,7) related to cyclic, bipartite volcanic eruptions. (see also Chapt. 7.4). An explosive, ?pulsating eruption column inaugurated the depositional cycle by producing the series of thin surging pyroclastic flows (siltstones); coarser-grained volcanic ash (sand-grade) usually characterised later eruptions. These pyroclastics were initially transported within high-density turbidity currents, with storm processes increasingly assuming responsibility for resedimentation of volcanic ash during the tectonic subsidence following the volcanic eruption. Large blocks of pumiceous tuff within more substantial channels might represent

accidental clasts of unwelded PHT incorporated *en route*.

The regressive epiclastic sequences, in contrast, would reflect "normal" progradation of small nearshore depositional systems during tectonically "quiet" periods. As volcanotectonic uplift, of equal magnitude to post-eruption subsidence, should precede the rhyolitic volcanic eruptions, it is intriguing that the regressive epiclastic and transgressive volcaniclastic sequences appear to be genetically distinct (*ie.* no coarsening-upward epiclastic sequences occur beneath primary pyroclastic units). The only way to resolve this conundrum is to invoke two separate sources for epiclastic and pyroclastic successions (Fig. 5.50). Palaeocurrents within shoreface successions at Pont-y-Gromlech indicate transport of considerable detritus by longshore currents; the absence of the rhyolitic "tuffites" within epiclastic sequences at Pont-y-gromlech and their insignificant thickness at Llyn Clyd (SH 635 597; Fig. 5.50) implies: 1) a volcanic vent for the pyroclastics along either the Idwal, or Glyder Fawr faults and/or 2) depositional segregation of the Llanberis Pass and Cwm Cneifion areas. Although the Snowdon fault block likely "bobbed up and down" in empathy with the tectonic changes observed at Cwm Cneifion/Llyn Bochlwyd, little sediment of any composition could be derived as the pervasively welded PHT formed a resistant and impervious "blanket" over potential sediment sources. The unwelded top portion of the PHT had already been erosively removed during the first marine transgression (unit 1) and shoreface successions at Pont-y-Gromlech (Fig. 5.30) directly overly the eutaxitically welded central portion of the ash flow. The lack of coarsening-upward regressive sequences preceding volcanic events would be a consequence of the scarcity of coarse-grained sediment rather than an absence of pre-eruption tumescence. In fact, where the volcanic eruption only consisted of fine-grained silt-grade ash (unit 3) deposited as primary pyroclastic flows, one does not even witness the

post-eruption volcanotectonic subsidence in the depositional record.

The above patterns of deposition, with a subtle interplay between sediment supply, rhyolite volcanicity, and basin tectonics, is replicated in other portions of the UCEF, with variation on the theme provided by different rates of tectonic subsidence, local topographic relief along related faults, and the composition of magmatic eruptions. Similarities and contrasts in the pattern of volcaniclastic sedimentation at three additional localities will be discussed in relation to the above factors.

5.4.2.2 CAPEL HILL

At Capel Hill, the UCEF again contains a series of fossiliferous shallow marine sediments separated by several rhyolitic tuff horizons (Fig. 5.51). Sandstones are more prevalent than within successions at Cwm Cneifion/Llyn Bochlwyd, and appear to be entirely epiclastic in origin. This higher "background" supply of sand enabled development of thin coarsening-upward sequences beneath the pyroclastic units in addition to the overlying better defined transgressive sequences. The common presence of amalgamated sets of hummocky cross-stratification thoughout again suggests an inner shelf setting above storm wave base. Evidence for deposition within reach of fairweather currents (probably waves) is provided by trough and tabular cross-stratification and non-graded, tabular, coquinas of brachiopods characteristic of *Dinorthis-macrocoelia* communities (Wright, 1979, p. 145) and isolated sets of vortex wave ripples which suggest water depths of less than 7 metres (Chapt. 5.5.1).

A variant on the depositional theme already established for the UCEF, is provided by a small shallow marine basaltic tuff cone (Capel Hill, *sensu stricto*, SH 723 581). As early as 1922, Williams recognised the phreatomagmatic (Surtseyan) origin of the deposit, attributing the upward increase in the proportion and size of microlithic lapilli and bombs to the relative amounts of magma and water in the volcanic conduit. Although

centroclinal dips now characterize the thickest hydroclastic deposits (see Howells *et al*, 1978, p.17-19; my Fig. 5.52), to the north of Capel Hill strata dip 17-20° toward the NW after removing regional dip, implying that the actual vent lay a short distance to the east of Capel Hill. Lithofacies within these marginal "foresets" include tabular beds of basaltic mudflows and high density turbidity currents, and low angle cross-stratification (Fig. 5.53); the "foresets" are a small-scale version of "foreset lithic breccia" unit of the Bedded Pyroclastic Formation at Cwm Glas (Fig. 5.54) which has been attributed to deposition on the margin of a Surtseyan-Strombolian tuff cone (P. Kokelaar, pers. comm., 1986). The only other monolithic primary basaltic volcaniclastics preserved as mappable deposits within Caradocian sediments subjacent to the Snowdon Volcanic Group, are several coarsening-upward hyaloclastic to pillow basalt/blocky lava sequences (*eg.* Fig. 5.49) directly beneath the Lower Rhyolitic Tuff on Gribin ridge (SH 644583 to 652594).

Basaltic tuffs also occur at Capel Hill in the form of several thin heterolithic sediment gravity flows. The uppermost package of flows (Fig. 5.55) represents the last input of coarse clastic sediment into this part of the basin (Fig. 5.51), and can be traced for more than 1700 metres parallel to flow before "disappearing" into offshore mudstones. The depositional unit is bipartite with a lower mudflow containing 1-2 cm clasts separated from an upper inversely graded blocky debris flow by a fine to medium-grained sandstones with upper flow regime plane bed lamination (Fig. 5.55) or siliceous siltstones (sections 9/10). These intervening lithologies are often incorporated within the upper debris flow as rafts of plastically-deformed sediment with proximal deposits and lithic blocks within more distal successions. Clasts include all subjacent lithologies, with angular clasts of fine-grained rhyolitic tuff (Fig. 5.55, photo) being particularly common in proximal deposits and lithic blocks of

pumiceous basalt, sandstone and siltstone increasing in abundance distally. The flow displays little variation in thickness over the first 1000 metres (sections 1-11), is continually inversely graded, and an exponential decrease in largest clast size (Fig. 5.55, graph) represents the dominant variation in facies over this distance. Further to the north, however, the lithic proportion drops off rapidly (sections 12-15), occasional large (ca. 150 cm) clasts are occasionally found in the base of the sediment gravity flow, and the deposit thickness increases from 3 to 6 metres. The rapid transformation from a viscous, cohesive, matrix-supported to a more dilute, cohesionless, debris flow (*cf.* Nemec & Steel, 1984) was probably a function of a change in topographic slope within the basin (see Fisher, 1983). With incorporation of substantial quantities of adjacent mud (and contained water), the flow was unable to support and/or transport its remaining large clasts.

5.4.1.2 BWLCH TRIMARCHOG

At Bwlch Trimarchog (SH 708 627), the UCEF highlights the "mud problem". The middle portion of the formation (Fig. 5.43, Units B-I) comprises a thick (ca. 400 m) succession of intercalated sandstone, siltstone, and mudstone. In the lowermost sandy unit (Fig. 5.43, enlargement) internal sedimentary structures within coarser lithologies include undulating horizontal lamination, low-angle cross stratification (Fig. 5.8), amalgamated sets of hummocky cross-stratification and shallow-moderate angle trough cross-stratification. Wave and current ripples occasionally mould the top of sandstone units. This assemblage of lithofacies again suggests deposition on a series of inner shelf sand bars and depositional lobes, with shoaling occasionally placing the sediment-water interface within reach of fairweather currents. The total absence of any organisation (*ie.* cyclicity, recognisable sequences) within nearshore successions and continual presence of intercalated mudstone beds throughout

implies a limited sand supply. Further testimony to the validity of the above line of reasoning is provided when one considers the entire depositional succession (Fig. 5.43). The lowermost sandstones, which are thought to have been deposited in 10-30 metres of water, abruptly overly 600 metres of cleaved mudstones which display no apparent vertical changes in grain-size. Although there is obviously more happening within the mudstones than "meets the eye", it is envisaged that all mud within the lower and middle portions of the UCEF was deposited in much less than 100 m of water. Muddy facies signify an absence of sand rather than deep environments. It is pertinent to note, in this regard, that no graptolite fauna have been collected from any mudstone or siltstone from the entire 50-1400 metre thick UCEF in North Wales.

5.4.2.4 LLYN COWLYD

The middle portion (Y) of the UCEF at Llyn Cowlyd (Fig. 5.56) has a slightly finer grain size suggesting a further distance from source terrains. Lithofacies within the fossiliferous and tuffaceous sandstones of the Multiplicata and Arkosic units (terminology from Diggens & Romano, 1968), for example, imply deposition within swash bars and shallow marine sand bars respectively.

Lithofacies associations displayed by the upper portion (Z) of the UCEF, however, are a unique. Here, heterolithic, inversely and normally graded conglomeratic sediment gravity flows are a common facies at the base of a fining-upward transgressive 100 m thick sequence. The type of sediment gravity flow changes upward with tabular, inversely graded, debris flows (Fig. 5.57) being replaced by distal channelised, normally-graded, fluidal flows (*ie.* high density turbidity currents; Fig. 5.58) at higher stratigraphic levels. Clasts, which include flow-banded and structureless rhyolite, tuffaceous siltstone, and quartzose sandstone, decrease in maximum size and abundance upwards in tandem with the above textural

changes, indicating an associated decrease in flow competence. These transformations in flow behaviour, in analogy with alluvial systems, would largely reflect a progressive decrease in depositional slope and available sediment on the submarine "fan" (*cf.* Nemec & Steel, 1984; Surlyk, 1984).

The insignificance of these conglomeratic lithofacies at Bwlch Trimarchog and their total absence at Capel Hill implies tectonic activity along the Llyn Cowlyd fault (Fig. 5.43, 5.60) with downthrow to the NW. The proximal debris flows would reflect an initial abundant supply of sediment in response to fault movement. With a progressive reduction in topographic relief along the Llyn Cowlyd "escarpment", sediment gravity flows not only became more dilute but rapidly diminished in importance as a mode of sediment transport on the submarine slope. Their distribution over a stratigraphic thickness of 40 m provides a crude estimate of the initial relief produced along the Llyn Cowlyd fault. The upwards increase in the abundance of storm-related bedforms (*eg.* 5.59) preserved within the succession (Fig. 5.56) at the expense of sediment gravity flow deposits signifies a return to the "normal" processes of sediment dispersal which characterize the UCEF. The absence of high angle cross-stratification throughout suggests deposition below fairweather wave-base.

5.4.1.5 DEPOSITIONAL HISTORY OF UCEF

Correlation of depositional sequences with the UCEF to reconstruct the depositional history of the formation is frustrating. It is enticing, at first, to try to correlate pyroclastic units across the basin. Thickness, grain size, and facies variations within the rhyolitic tuffs, however, indicate that two distinct suites of subaqueous pyroclastic flows are present, probably derived from opposing eruptions on the Idwal and Llyn Cowlyd faults (Fig. 5.43). Although there is not enough chemical variation in trace elements to statistically discriminate between between the pyroclastic units at the various localities (Chapt. 2.2.4.4), correlation

on each side of the Tryfan depocentre (Fig. 5.43, units 1-8, and A-H) can be achieved through consideration of the stratigraphic position and facies of adjacent epiclastic successions.

On the NE side of the basin, the consistent occurence of faunal assemblages (unit C) dominated by *Dinorthis multiplicata* brachiopods (which are absent at other stratigraphic levels in the UCEF) within fine to medium-grained sandstones overlying the first "tuffite" horizon (unit B) confirm correlation of this pyroclastic unit. This Multiplicata-dominated marker horizon (Mutliplicata sandstone of Diggen & Romano, 1968; see Wright, 1979, p. 143-147, his unit M) can also surprisingly be identified in the lowermost portion of the UCEF at Cwm Cneifion (Fig. 5.44, unit 2; BGS PD/80/27 No. 6,7) and appears approximately 15 stratigraphic metres above the PHT (Appendix, section U2) in the condensed succession at Llyn Clyd (BGS faunal locality PD/80/27 No. 1). If these basin-wide correlations are correct, than this time period represents a unique episode in basin history when a prolonged period of tectonic quiescence and little differential subsidence, allowed proliferation of an opportunistic species over an area of at least 100 km^2. It remains to be determined whether this palaeontological and tectonic "event" can be identified in other parts of the Welsh basin. The sandy middle-upper Allt Ddu or Carn y Parc formations represent obvious candidates for examination near Betws-y-Coed and Bala (Campbell, 1983). It is interesting to note that unlike most other brachiopod or trilobite species, *Dinorthis multiplicata* appears to be stratigraphically restricted to only a few faunal horizons (middle assemblage zone of Cwm Rhiwarth Fm. and upper part of Teirw Siltstone Fm.) within Caradocian successions in the Berwyn Hills (Bancroft, 1945; Brenchley, 1978, p.160). What is even more intriguing in North Wales, is that at Cwm Cneifion/Llyn Idwal the Multiplicata sandstone is also underlain by a fine-grained rhyolitic ash flow tuff (Unit 3). The

synchronicity of volcanic eruptions from separate vents on the Idwal and Llyn Cowlyd faults must be related to tectonic stress patterns within the continental crust.

The sandstone dominated successions overlying the Multiplicata sandstone, in contrast, are virtually impossible to correlate across the basin with available data. One would intuitively suspect that the rhyolitic tuffs and sandstone lobes from the NE and SW sides of the basin were also synchronous in their development. More palaeocurrent and geochemical information is necessary before the thin depositional systems can be assembled at a reasonable level of precision to test this hypothesis. The synopsis of the depositional history of the UCEF presented in Figure 5.60 (see caption) thus only approaches regional considerations at the broadest of levels. The pattern of deposition found in North Wales is broadly replicated in other portions of the Welsh basin. Most uppermost Soudleyan successions (middle portion (Y) of UCEF), for example, were deposited well above storm wave base; dessication cracks have been reported from the correlative uppermost Allt Ddu Fm. near LLanuwchllyn (Lockley, 1980, p.31). A picture emerges of a shallow epicontinental sea with a limited supply of sand being reworked into muddy shoals and sand bars which were occasionally emergent. In contrast, during lowermost Lower Longvillian (or Woolstonian) times, the basin gradually subsided below storm wave base (upper portion (Z) of UCEF; see also Brenchley, 1969, 1978) except where buoyant magmas supported fault blocks (Chapt. 6).

5.4.2 EVIDENCE FOR TIDES

Although tidal deposits are not unknown from ancient epicontinental seas (*eg.* Daily *et al*, 1980), and on modern shelves there appears to be a positive correlation between shelf width, tidal amplitude and tidal current velocities (Klein & Ryder, 1978), their identification has remained elusive within the Welsh basin. Sandwave deposits from the Arenig Garth grit on St.

Tudwals Peninsula (G. Tegardine, pers. comm., 1987) are one of the few indisputable Lower Palaeozoic examples of tidal processes. Evidence is presented below, however, to suggest that in certain depositional situations, tides were as important as storm processes in the distribution of shallow marine sediments.

A) Vuggy Mbr., Ddeugwm Fm., Tryfan anticline.

The Vuggy Mbr. comprises a 3-8 m thick succession of trough cross-stratified medium-grained sandstone of homogenous composition interpreted to represent a tidally-influenced sandwave complex (Fig.5.62). The succession overlies a series of thinly bedded, normally graded subaqueous pyroclastic flows (Tuffite Mbr.) and the quartzofeldspathic composition of the Vuggy Mbr. (Fig. 3.111) suggests a genetic association with the subjacent pyroclastic flows; the member would represent reworking of coarse-grained volcanic ash from the second part of the volcanic eruption (see Chapt. 7.4). The sandwave complex is divided into three depositional units (Fig. 5.62) by several westward dipping, low-angle 5-10 cm siliceous mudstone horizons (Fig. 3.21a). These large scale, low relief surfaces (in effect master bedding planes) would reflect fluctuations in the rate and direction of bedform migration. Palaeocurrents within each unit are bimodal, with the dominant direction of cross-strata dip varying between the two units. Cross strata varies in set size from 20-90 cm, with the mean set thickness decreasing upwards. The overall direction of bedform migration, as shown by the composite palaeocurrent rose, was towards the north, perhaps reflecting the additional influence of currents related to the high-energy storm waves that often came from the south (Chapt. 5.5.1). Similar to other ancient sandwave complexes (Roda sandstone; Nio, 1976), the Vuggy Mbr. was deposited in a tectonically unstable, rapidly subsiding basin, and is overlain by bioturbated offshore marine sediments. Correlative successions at Craig yr Ysfa (Fig. 5.62, see Appendix, section

F1, 0-10 m), where the entire Ddeugwm Fm. is only 15 m thick (cf. 180 m to the SW of the A5) and overlies a high-level ?extrusive rhyolitic pluton, are envisaged to have been deposited in a tidal flat/shoreface setting. This is implied from tangentially-based trough cross-stratification with numerous reactivation surfaces (?mud drapes) and linguoid current ripples yielding bimodal (east-west) palaeocurrents, and flaser bedding.

B) Upper Cwm Eigiau Fm. Pont-y-Gromlech.

A fining-upward sandstone sequence directly beneath the Lower Rhyolitic tuff is interpreted as representing a transgressive non-barred shoreline (Orton, 1986). Although lithofacies indicate that storm and wave processes were dominant, several occurrences of herringbone cross-strata (Fig. 5.31) suggest that tidal-currents might have been operating. In addition it is suspected that the fairweather currents(s) responsible for the shore-parallel migration of the large sandwave complexes in both this and the subjacent shoreface successions at Pont-y-Gromlech included a tidal component.

If the above interpretations are correct, than tidal processes were more important in the distribution of sediment within Caradocian shallow marine environments in North Wales than has previously been suspected. The fairweather shore-parallel currents within the Middle Racks Mbr. (Chapt. 5.3.3) and Lower Racks Mbr. mudflat/barrier island complex are additional candidates for tidally-influenced depositional settings.

The preservation of tidal bedforms/facies in the depositional record in North Wales is clearly a function of the amount and grain size of sediment supplied. During periods of rapid rates of sedimentation and/or coarse grained detritus, the higher preservation potential of storm deposits, and increased importance of undercurrents from fluvial discharge, combined to reduce the apparent significance of tides within the Welsh basin. As tidal ranges were higher in the Ordovician due to the closer

proximity of the moon to the earth, even microtidal conditions (ca. less than 2 m) might result in shoreline fluctuations of almost 1 km on an "typical" braidplain-delta (gradient of *ca.* 3 m/km *cf.* Skeidararsandur coastal plain). This has direct implications for entry of ash flow tuffs into the sea as volcanic eruptions will preferentially occur during maxima of semi-diurnal tides (Golombek & Carr, 1978).

5.5 THE WELSH BASIN: A STORM-DOMINATED, MICROTIDAL, EPICONTINENTAL SEA.

5.5.1 SHALLOW MARINE PROCESSES & BATHYMETRY

The above analysis indicates that both fairweather and storm waves were the dominate depositional processes responsible for the distribution of nearshore and shelf sediments. Tidal currents occasionally expressed themselves during periods of reduced sediment supply or on low-gradient coastal plains; sediment gravity flow were periodically important during volcanic events or associated with local topographic relief along fault scarps.

The shallow marine processes and inferred depositional environments can first be used to provide broad constraints on the bathymetric gradient. On modern coastlines, fairweather wave base varies according to the flux of wave energy within the basin with values ranging from 20 m on the high-energy Oregon coast to 5 m adjacent to Georgia. In this light, it would be reasonable to assume an intermediate depth of 10-15 metres for an epicontinental sea such as the Welsh basin. Identification of storm wave base is more difficult in depositional sequences as small-scale rippling within fine-grained facies is usually not discernible. Some limit on water depths within offshore successions, however, is provided by both hummocky cross-stratification (HCS) and wave ripples. Harms (1975) used vertical thickness of deposits containing hummocky cross-stratification in progradational sequences to estimate maximum water depths of 30 m; comparable depths are obtained in North Wales from deltaic successions.

Although this is not the place for comprehensive palaeohydraulic interpretations, calculation of wave parameters from available sand and gravel ripples using linear wave theory (*eg.* Clifton & Dingler, 1984; Leckie & Duke, 1984; Diem, 1985) also identify useful limiting conditions on both water depth and wave climate. All field data collected from wave ripples, as well as calculated parameters (using Diem, 1985) are tabulated in Fig. 5.63. Although absolute values derived from application of linear wave theory must be treated with extreme caution due to 1) tectonic deformation of ripple profiles and 2) uncertainty whether ripples were orbital types (see discussion in Appendix 4), the values obtained confirm suspicions from empirical analysis of depositional facies (Chapt. 4.4.1, 5.3.5). The maximum wave periods (8.2 secs) were derived from coarse-grained nearshore wave ripples of the Capel Curig braidplain delta. Assuming that this represents the most violent storms in the Welsh Basin, then the the maximum depth at which storm waves could exceed the threshold velocity to set fine sand in motion (0.127 m/sec; Bourgeois, 1980) would be *ca.* 60 metres.

5.5.2 DISTRIBUTION OF WAVE ENERGY & RELATED WATER CIRCULATION

Calculations using linear wave theory also confirm suggestions from depositional facies in deltaic successions (Chapt. 4.4.1) that wave power approaching from SSW to SSE was an order of magnitude larger than wave energy approaching from any other compass direction (Fig. 5.64). The data set is obviously biased by the dominance of examples from the NE-SW orientated delta front of the Capel Curig depositional system (*eg.* Fig. 4.54). Arguments that the distribution of wave energy is merely a ficticous manifestation of wave refraction related to coastline orientation can be easily dismissed, however, as large wave ripples also characterize abandonment facies of the Bochlwyd delta (Fig. 4.78, section E9, 39 m) where they migrated parallel to the inferred NW-SE (*ie.* Llanberis pass

fault) coastline. In addition, all offshore wave ripples (*eg.* at Bwlch Trimarchog or Dolwyddelan) indicate similar directions of wave oscillation.

The pattern of sediment movement resulting from surge and retreat of such waves against a coastline will obviously be influenced by its orientation relative to maximum wave energy flux. The shore-parallel drift of sediment to the NE during progradation of the Capel Curig Mbr., and Racks Mbr., and dispersal towards the N along the Upper Cwm Eigiau Fm. shorelines at Pont-Y-Gromlech could be explained by this control. The patterns of sediment dispersal, however, are not as simple as this as there are numerous deltaic systems where alongshore currents flowed the "wrong" way (*eg.* during retreat of Capel Curig Mbr., longshore currents flowed at SE along a NW-SE coastline; Gwern Gof Mbr.). As sandy sediment was preferrentially dispersed to the left of river mouths (where it can be inferred), it is more enticing to appeal to geostrophic currents to account for the assymetric distribution of sediment in coastal depositional systems. The Coriolos force during the Ordovician would also be significantly higher than that today due to the more rapid rotation of the earth; in the southern hemisphere, storm-currents retreating from a shoreface or mouth bar would preferrentially flow to the left.

5.5.3 BASIN PARAMETERS

Finally, the ancient storm wave periods of 5-8 seconds for waves arriving from the south implies an association with extratropical cyclones (gales) or winter storms rather than hurricanes. The clustering of wave approach directions (and hence formative winds) also argues against a hurricane origin for the storms. As palaeomagnetic data suggests that Wales (Avalonia) was situated at a latitude of 20-30° S in the Caradocian (Piper, 1978), and indicate a net clockwise rotation of the British Isles after closure of the Iapetus (Smith *et al*, 1980, map 73), enhancement of SE trade winds appear to be the most viable explanation for genesis of the

storm waves. Not surprisingly, the continual arrival of winds from the south influenced the distribution of pyroclastic airfall tuffs in North Wales. Ash beds within deltaic sequences in the Tryfan anticline (*eg.* Fig. 4.3, 4.6, section C2,C5) thin to the north; the thickness of tuff horizons within the 4th volcanic mbr. of the Capel Curig Volcanic Fm. containing accretionary lapilli is an order of magnitude greater within sequences to the north of the vent (Howells & Leveridge, 1980) at Glyder Fach (Chapt. 7.2.4).

Further support for the above wave pattern arises when one considers the wave energy regime which influenced nearshore sedimentation during other time periods during the Lower Palaeozoic. For example, large gravel ripples (60 cm wavelength) from Cambrian strata at Cwm Graianog near Bethesda, North Wales (Crimes, 1970, his Fig. 21,24,25) indicate approach of storm waves from the SE obliquely towards an inferred WNW-ESE oriented coastline (my interpretation of palaeocurrent data in Fig. 26 for coastline). Wave ripples from Arenig successions on Lleyn (Fig. 5.66) have similar crestline orientations, and are thought to reflect advance of storm waves from the SSE towards a NE-SW orientated shoreline (G. Tegardine, pers. comm., 1987).

The consistent approach of maximum wave energy from the SE to S, regardless of coastline orientation, throughout 80 million years of geologic time has rather profound ramifications for plate tectonic and paleogeographical reconstructions that one erects for the Welsh Basin during the Lower Palaeozoic. In the first place, it certainly negates substantial northward drift of Avalonia from 60 to 20°S during Cambrian to Silurian times as postulated by some authors (*eg.* Cocks & Fortey, 1982) based largely on the absence of carbonate reefs from the south of Britain; winds would change from westerlies to SE trades over this latitude. It also implies little effective tectonic rotation of Avalonia **during** the

Lower Palaeozoic. On a more local scale, if the suggestion that the largest storm waves approached from the SE is correct, it leads to radically different interpretations of palaeogeography in other parts of the Welsh basin. In an analysis of the environment of deposition of the Woolstonian (Caradoc) Alternata Limestone of East Shropshire, Hurst (1976) related a "mosaic" of storm-influenced nearshore environments to an assumed fault-bound English Midlands landmass further to the SE (Fig. 5.67a). It should be noted that outcrop parallels the depositional strike and no palaeocurrent data were presented to support the reconstruction. This palaeogeography was later adopted (Brenchley & Newall, 1982) for younger (Marshbrookian) storm-dominated sandstone lobes at the same locality with the only palaeocurrent data presented being an inferred shore-parallel orientation of wave ripple crestlines (Fig. 5.67b). In light of the evidence presented above regarding wave conditions in North Wales, does it not seem more reasonable to relate depositional environments to an uplifted horst block to the **NORTHWEST** (Fig. 5.67c) bound by the known Church Stretton and Pontesford faults.? The segregation of the Shelve and Shropshire depositional basins would also account for their contrasting volcanological and sedimentological patterns of deposition. Some additional evidence will be tabulated later (Fig. 8.3) to suggest that the Welsh basin **never** had a NW and SE margin (Anglesey-Rosslare horst and Church Stretton faults respectively) but was in fact a NORTHWEST-SOUTHEAST oriented back-arc basin related to subduction of the Tornquist sea.

CHAPT. 6. **BASIN EVOLUTION: THE UPS AND DOWNS**

> "we cannot doubt that during the long eruptive period of Llandeilo and Caradoc times, the Snowdon volcanic field, like almost every modern area of volcanism, was one of repeated disturbances, of faulting and tilting, due to magmatic movements at depth. These suggestions are emphasisized in the belief that the region was already considerably fractured and plicated before the onset of the main Siluro-Devonian orogenic movements."
>
> *Williams & Bulman, 1931.*

6.1 INTRODUCTION

In modern volcanic fields, uplift due to buoyant forces associated with the emplacement and vesiculation of magma is usually seen to occur before volcanic eruptions with subsidence post-eruption (Fig. 6.1). Topographic changes have been determined using tilt and levelling measurements and photogrammetric techniques (*eg.* Moore & Albee,1981; Lipman *et al*, 1981). The amount of uplift/subsidence varies according to the tectonic regime, nature of eruption, size of eruption, and the type of magma involved. The control of the last variable is well illustrated by a comparison of uplift before the 1967 eruption of Kilauea and the May, 1980 eruption of Mt. St. Helens, two eruptions of similar volume). All of these observed eruptions have been ?fortunately for mankind quite small involving magma volumes of less than $0.2km^3$.

In the geological record, pyroclastic eruptions, especially those associated with large calderas, can have magma volumes exceeding $1000km^3$ (Lipman, 1984). These large caldera complexes all display similar structural features associated with a common evolutionary trend (Smith & Bailey, 1968; Fig. 6.2). Premonitory activity includes precaldera volcanism, associated tectonism and broad doming that records the rise and fractionation of a shallow crustal batholith. Main caldera forming events are triggered by voluminous eruptions of silicic pyroclastic material. Evacuation of the magma chamber results in both caldera collapse and

regional detumescence (3: FIG. 6.2). The caldera fill is dominated by thick ash flow deposits that were locally contained by subsidence of the caldera floor concurrently with the volcanic eruption (see Lipman, 1984). Postcollapse activity often includes resurgent uplift of the caldera related to the renewed rise of the subjacent magma body.

A caldera, by definition, must show evidence for subsidence of a fault block associated with volcanism. Numerous calderas have been resurgently uplifted. There is even evidence at several young calderas that uplift is still under way; ground deformation studies have documented significant recent tumescence at both Long Valley and Yellowstone calderas and in the Rio Grande rift near Socorro (Savage & Clark, 1982; Reilinger *et al*, 1977; Reilinger & Oliver. 1980). Geophysical data often indicate the presence of mid-crustal magma bodies (Denlinger & Riley, 1984) in these areas.

In the caldera cycle, it is the evidence for the INITIAL broad doming and inflation related to the upward movement and accumulation of large bodies of silicic magma that is most elusive. Lipman (1984) in his review of Tertiary calderas in the Western United States identified only two calderas which have geological structures which record premonitory uplift: the Timber Mountain Caldera in Nevada (Christiansen *et al*, 1977) and the Trans-Pecos Chiantini caldera (Henry & Price, 1984). This is surprising when one considers that the same mechanism can produce substantial resurgent uplift over a large area, an example being the 600 metres of resurgent tumescsence over an area of 2500km^2 at the Toba volcanotectonic depression in Sumatra. Much evidence for regional tumescsence can be destroyed during formation of calderas, and in volcanic fields with little informative epiclastic sedimentation associated with the volcanicity one is forced to use geometric relations such as the tilt of country rocks with respect to caldera structures to deduce that tumescsence pre-dated caldera forming eruptions. Even this information may be lost if structural

downsagging (Walker, 1984) occurs before formation of caldera structures such as ring faults (see Rytuba & McKee, 1984).

In North Wales, volcanotectonic uplift/subsidence has long been suggested solely to account for occurrence of welded ash flow tuffs within sequences presumed to be largely marine (*eg.* Rast, 1961; Fitch, 1967; Beavon, 1974). In the most elaborate development of this idea, Beavon (1980) related rhyolite and basalt volcanism to cyclic episodes of emergence and subsidence of volcanic islands. More recently, it has been in vogue to consider that welding within these ash flow tuffs took place under water (Francis & Howells, 1973; Howells & Leveridge, 1980); although these authors recognised the marine transgressions which occur after volcanic eruptions, they failed to consider that matching depositional or tectonic regressions might precede tuff emplacement, and assumed that environmental conditions within the basin did not change with respect to volcanic events.

In this chapter, changes in depositional enviroments, style of sedimentation and basin paleogeography will be considered with respect to laterally extensive ash flow sheets or volcanic units. These are, with minimum depositional volumes, as follows:

Gwern Gof Tuff		2 km^3
Capel Curig Volcanic Fm:	Garth tuff	28 km^3
	Racks tuff	15 km^3
Pitts Head Tuff		70 km^3
Lower Rhyolitic Tuff		54 km^3

The location of ash flow tuffs considered in the stratigraphic column are indicated in Fig. 6.3. The point of this chapter is not to reiterate ideas introduced in previous portions of the thesis; through a brief synthesis of the successive changes in basin configuration and palaeogeography, however, it can be shown that volcaniclastic sedimentation was strongly controlled by volcanotectonic uplift and subsidence. The amounts of pre-eruption tumescence and post-eruption detumescence that one sees both within North

Wales and generally within the geological record will be later related to the strength of the continental crust.

6.2 BASIN ANALYSIS

Palaeogeographical slices through basin history will be examined in ascending stratigraphical order; a crude estimate of absolute time is provided in Fig. 6.3 by assuming (falsely) constant rates of epiclastic deposition through the Costonian to Longvillian for the Tryfan depocentre.

6.2.1: ASSOCIATED WITH THE GWERN GOF TUFF

The Gwern Gof tuff (GGT), a largely unwelded garnetiferous ash has only been conclusively identified in the Tryfan anticline over an area of less than 10 km^3, despite its substantial thickness (40-60 m) and distinctive lithology. A garnetiferous schist described by Williams (1927) from Llanberis Pass is one possible candidate for further consideration. A source to the ESE (Llyn Cowlyd fault?) is postulated based on depositional slope indicated by subjacent deltaic sequences (Fig. 6.4a). The sediment supply to these prograding deltaic systems was interrupted a short time (*ca.* 20,000 years) before eruption of GGT. The top 10 m of the Gwern Gof Mbr. in southern portion of Tryfan anticline comprises a coarsening-upward sequence of largely bioturbated, fine- to medium-grained, locally magnetic sandstone displaying trough and low-angle XS, with rare sets of undulating lamination (Appendix, section C4, Fig. 4.3). Facies are thought to represent deposition within low-moderate wave energy shorefaces.

The continuance of shallow marine sedimentation after emplacement of the Gwern Gof Tuff implies subsidence of at least 50-60 metres associated with the eruption. More detailed palaeogeographical reconstructions directly above the ash flow are difficult to determine. The greater thickness of nearshore reworked facies on the west side of the anticline (Appendix, section C1, 0-10 m) hints that depositional "polarity" within the basin might have "flipped" in conjunction with the volcanic event.

Definitive evidence for a completely different basin palaeogeography and patterns of sediment dispersal, and by inferrence tilt of underlying fault blocks, is provided by the Tryfan Fach braidplain delta (Chapt. 4.3.2) which prograded to the SE.

6.2.2: ASSOCIATED WITH GARTH TUFF

The Garth tuff is the lowermost and most extensive volcanic unit of the Capel Curig Volcanic Formation (CCVF) and extends as an ash flow of various form (Howells & Leveridge, 1980) from Llandudno in the north to Dolwyddelan and Bala in the SE. A source to the north has long been inferred (Howells *et al*, 1979) from both distribution of tuff facies and interpretations of depositional environments. A more specific vent location along the Conwy fault is proposed here (Fig. 6.9a) to better integrate palaeocurrents from Craig yr Ysfa, the marine environments at Conwy, and aspects of tuff along the west side of the Idwal syncline. The eruption and emplacement of the ash flow coincided with the climax of coarse epiclastic deposition in northern Snowdonia as represented by the Moel Siabod Mbr.. Depositional facies and related environments within this stratigraphic unit tend to mirror those within earlier depositional systems, with alluvial fan, braidplain, and shallow marine environments all recognised.

At Bwlch-y-Ddeufaen (App. Section S2), Llyn Dulyn (App. Section T1), and Craig yr Ysfa (Fig. 6.5, 6.9a), facies associations indicate deposition in stream dominated alluvial fans, supporting earlier interpretations by Howells and Leveridge (1980). Available palaeocurrents imply uplift to the east and sources (fan apices) along the Conwy fault is envisaged. Sandstone petrography (Chapt. 3.4.2) indicates derivation of most sediment from epiclastic reworking of quartzose volcaniclastics with the proportion of material derived from rhyolitic volcanics increasing to the north; there is no evidence of concurent pyroclastic activity. The region between the

Melynllyn and Afon Ddu faults (Fig. 6.5, 6.9a) was structurally uplifted for most of the Tryfan Fm. time period and successions at Llyn Dulyn (Appendix, section T) are stratigraphically condensed.

To the south of Craig yr Ysfa in the Tryfan anticline, the Moel Siabod Mbr. comprises a coarsening-upward succession of siltstone to medium-grained sandstone. The top portion consists almost entirely of sandstone, with trough cross-stratification the dominant sedimentary structure. Palaeocurrents are unidirectional toward the SSE. Numerous poorly defined sequences, defined by upward decreases in both set size and grain size can be recognised. They are separated by either fine-grained siltstone or very fine-grained sandstone displaying small current ripples, horizontal lamination, and rare desiccation cracks (section C3, xxm; Howells & Leveridge, 1980, p.12). It is difficult to understand the basis for the repeated contention by previous researchers (*eg.* Howells *et al*, 1979, p.616 and subsequent publications) that ash flow tuffs transgressed into "shallow water environments south of Craig yr Ysfa"; the most recent interpretations (Kokelaar *et al*, 1984, p. 314-315) favour a marine deltaic environment which was of sufficient depth to allow the entire ash flow (presently *ca.* 80m thick) to be subaqueously emplaced. Even the palaeogeographic maps of these authors (Howells & Leveridge, 1980, their Fig. 3vi) correctly allude to the common presence of fluvial and/or lacustrine environments in the Tryfan antiform. Although the author agrees that a wet substrate is necessary to allow formation of the irregular loaded base to the Garth Tuff (Francis & Howells, 1973; see Chapt 7.2.2), this condition is not a prerequisite for subaqueous environments. A similar deformation pattern of comparable scale has been described from loading of basalt onto water saturated fluvial/floodbasin sands from Middle Proterozoic successions in Australia (Needham, 1978); although there was no evidence that minimum fluidisation velocity of subjacent sediments was exceeded (cf. Kokelaar,

1979), the flow of steam through the sandstone under pressure was still attributed a major role in decreasing the viscosity of the sediment and facilitating rheological deformation.

By analogy with previously deposited successions in the Tryfan anticline, sediments in the upper portion of the Moel Siabod Mbr. are thought to reflect deposition within fluvial channels on a broad braidplain. The repetitive lateral migration of channels would produce the tabular sand sheets observed. Although the thin floodbasin sequences indicate some "standing water" (*ie.* centimetres), there is no sedimentary evidence within any depositional successions for the existence of deeper (*ie.* metres), more permanent lacustrine environments. The slightly finer grain size which characterizes the uppermost 20-30 m of fluvial successions is attributed to a subtle change in source region parameters, such as back-stepping of the basin margin, immediately before tuff emplacement. It is not necessary to invoke a change in depositional environments.

In the Capel Curig anticline, the depositional pattern does not appear to be any different (Fig. 6.5, 6.9a). The lowermost coarsening-upwards succession would again reflect progradation of a braidplain delta; the seaward-extending wedge of sandstone from the previously deposited Capel Curig delta system would account for the shallow water depths (inferred from sequence thickness) into which the Moel Siabod depositional system prograded. Depositional facies suggest a low wave energy relative to fluvial input on this delta front. Fluvial sediments capping the deltaic sequence display stacked-sets of trough and planar cross-stratification (Fig. 6.6) which yield unimodal palaeocurents to the SE (Fig. 6.5). Although sandstones immediately beneath the Garth tuff are extremely difficult to interpret because of poor exposure and liquefaction of substantial thicknesses (5-15m) of sediment associated with tuff emplacement, a distal braidplain environment is still envisaged. This is

based on: 1) little upward change in grain size, 2) common presence of magnetic heavy minerals, 3) total absence of fossils, and 4) all sedimentary structures observed were trough cross-strata. The shoreline immediately beneath the Garth tuff is envisaged to have been in the vicinity of Moel Siabod (hence the member name); the southern limit of the "transitional" environment depicted by Howells *et al* (1985) would be a good approximation.

The deposition of the Moel Siabod Mbr. represents a continuation of the pattern of sedimentation established by the previous members of the Tryfan Fm. (Fig. 6.9a); each braidplain system carried finer-grained sediment than its predecessor and methodically stepped further toward the south before encroaching upon marine environments. All of the braidplains were structurally bound by the uplifted Llanberis Pass fault system, which directly contributed rhyolitic pyroclastics to the Tryfan Fach and Capel Curig Mbrs.. Although this highland area along Llanberis pass was smaller than during earlier depositional episodes of the Tryfan Fm., it was still sufficient to deflect both the Garth and later Racks tuffs to the SE after they entered the Tryfan depocentre (Fig. 7.6, 7.15). By comparison with modern alluvial fans (*eg.* Boothroyd & Ashley, 1975), *ca.* 25 m/km would represent a liberal estimate of fan gradient; this implies *ca.* 250 m of relief between braidplain environments and alluvial sources along the Conwy fault, and assuming comparable altitudes of the tops of basin margin fault blocks, also along Llanberis Pass.

The role played by the NW "margin" of the Tryfan depocentre braidplains is extremely difficult to ascertain. In the first place, the succession from Gwastadnant to Pen-yr-Ole Wen (Fig. 6.9a) is probably condensed. In an attempt to correlate Ordovician ironstone deposition in North Wales, the occurence of Llandeilo acritarchs with depositional units at a position close to Arenig sediments was used to imply breaks in

stratigraphy (Trythall *et al*, 1987, p.39) in Llanberis pass; later erosional unconformities have been demonstrated at this locality (Chapt. 4.3.7.3) by the removal of *ca.* 40 m of proximal mouth bar deposits from the Gwastadnant fan delta. Similar tectono-stratigraphic events are also envisaged to be present within correlative sediments of Tryfan Fm.. When deposition did occur, facies suggest fluvial or lacustrine environments. Sediment liberated during episodes of uplift would be fine-grained (*eg.* reworked Nant Ffrancon mudstones) and go unnoticed within either braidplain sequences or sandstone petrography. It is probably safe to assume uplift to the NW of Idwal fault immediately prior to emplacement of the Garth tuff.

In the Dolwyddelan syncline (Fig. 6.8), the Moel Siabod Mbr. consists of siltstone to medium-grained sandstone deposited in shallow marine environments. Near Cwm Fynhadog Isaf (SH 685 582), lithofacies such as low-angle and planar-tabular XS, numerous tabular coquina beds, and hummocky or swaley XS suggest deposition above storm wave base well within the reach of fairweather currents. Lateral variation in grain size and lithofacies indicate most sediment came from the WNW. This succession is stratigraphically condensed when compared with those further east; it is envisaged (Fig. 6.9a,b) that flexural interference of the Llanberis Pass & Cwm Bowlyd/Trawsfynydd fault trends prevented full basin development. The area is also problematic with respect to palaeontology (Tunnicliffe, BGS PD/85/213) with Longvillian fauna appearing to have developed much earlier than anywhere else in North Wales. Where a "full" basin did form to the east, depositional facies near Penmachno such as "micro-hummocky" cross-stratification (Fig. 6.7) indicate deposition in deeper water close to storm wave base (*ie. ca.* 40 m). Both deltaic sediment from the Moel Siabod Mbr. as well as the Garth tuff itself were preferrentially deflected towards this basin by the Cwm Fynhadog "bulge".

The emplacement of the Garth tuff obviously produced instantaneous changes in the palaeogeography described above, with the coastline immediately advancing *ca.* 5-8 km to the south (neglecting subsidence during emplacement). As a largely welded sheet of ash flow tuff, where deposited subaerially, it obliterated previous sources of sediment along the NE margins of the basin, and tended to level out pre-existing topography. The formation of a flat to low-gradient coastal plain and eventual marine shelf combined with a locally abundant supply of sediment from unlithied disrupted ash flow tuff was extremely favourable for building of barrier islands and sand bars by wave processes (Fig. 6.9c; Chapt. 5.3.2). These depositional features retreated landward (N) during the marine transgression after tuff emplacement. The amount of tectonic subsidence was greatest in the south with sequences to SE of Llyn Cowlyd fault (Chapt. 5.3.2) indicating as much as 70 metres of post-eruption detumescence.

6.2.3: ASSOCIATED WITH RACKS TUFF, 3rd and 4th VOLCANIC MBRS. of CCVF.

The Racks Tuff (2nd volcanic Mbr. of Capel Curig Volcanic Fm.) extends as a continuous sheet from its probable source on the Conwy fault near Cefn Maen Amor (Fig. 6.10) to Capel Curig, diminishing in thickness from 100 to 10 m over this distance (Howells & Leveridge, 1980). North of Craig yr Ysfa the tuff rests directly on the underlying Garth tuff. In the Capel Curig anticline, however, 80 m of epiclastic sediment separate the first two volcanic members. The Racks tuff is immediately underlain by unfossiliferous mudstones of probable lower shoreface environment (*ie.* less than 20 m water depth) thought to represent resedimentation of the "lagoon" successions which existed to the N. The total absence of any sand supply precludes full determination of the tectonic changes that probably did occur.

The Racks tuff is largely overlain by cleaved mudstones of uncertain environmental affinity. Along the SW margin of the Tryfan anticline,

however, fossiliferous medium-grained sandstone displaying low-angle and undulating cross-stratification passes upwards into thinly bedded (15-20cm), normally-graded siliceous siltstone (Appendix, section B3). Succession is thought to represent a transgressive, storm-dominated shoreface; lithologies and distribution of facies in the Capel Curig anticline indicate sediment represents redistribution of unlithified portions of the Racks tuff.

There is no sedimentary evidence for any tectonic changes associated with the eruption and emplacement of either the 3rd or 4th members of the CCVF. The former developed a low-relief subaerial volcanic field about Carnedd Llewelyn; the latter produced a small, moderate relief, volcanic pile which quickly emerged from shallow (20-40 m) water depths to form a largely subaerial tuff cone. The eruption of the CCVF represents the last subaerial volcanicicity in NE Snowdonia; after emplacement of the 3rd & 4th member tuffs, the area slowly subsided beneath the Welsh seas as volcanic activity switched to SW Snowdonia.

6.2.4: CORRELATION WITH SW SNOWDONIA

The total absence of any expression of LVG volcanicity about Moel Hebog and Moel Ddu makes correlation of successions between the Tryfan and Beddgelert depocentres extremely difficult (Enclosure 1, Fig. 6.9b). The lowest sand-dominated sedimentary succession within the Caradoc in SW Snowdonia is the 400-600m thick Prenteg Fm. (Portreuddyn slates & Prenteg Grits of Shackleton, 1959). As tectonic changes (direction of uplift or subsidence) appear to be synchronous throughout North Wales, it is strongly suspected that the Prenteg Fm. represents the time equivalent of the Llewelyn Volcanic Group in SW Snowdonia. Shelly fauna within its lowermost siltstones and the base of the overlying Moel Hebog Mbr. (Gorllwyn slates of Shackleton, 1959) suggest it spans a Harnagian-Soudleyan age (Harper in Shackleton, 1959). The formation comprises a monotonous, poorly defined,

series of interbedded silstones and very fine-grained sandstones derived from the WSW. At Yr Ogof (SH 570 460), lithofacies through the top 200 m of the member include horizontal bedding, low-angle and swaley cross-stratification, small sets of trough XS, and planar horizontal lamination (Appendix, Section L5). The continual deposition in a narrow bathymetric window between storm and fair weather wave base over a thick stratigraphic interval implies gradual basin subsidence in tandem with sedimentation, and little variation in source region characteristics. Occasional debris flow deposits containing clasts of porphyritic rhyolite would reflect rare tectonic activity in SW Snowdonia during this time period.

6.2.4: ASSOCIATED WITH PITTS HEAD TUFF

The Pitts Head Tuff Fm. has three main geographical expressions (Reedman *et al*, 1987; Enclosure 3), with two outflow tuffs in the NE correlated by geochemistry with a thick (*ca*. 700 m) succession of parataxitically welded ash flow tuff in the SW considered to represent an intracaldera fill (Roberts, 1969). North of Moel Hebog (Fig. 6.11), only the lowest outflow is present; sediments enveloping this pyroclastic unit are the subject of the discussion here. In SW Snowdonia, the bottom portion of the Lower Cwm Eigiau Fm. (Moel Hebog & Moel Ddu. Mbrs. respectively; Fig. 6.11a) represent progradation of fan and braid deltas as discussed earlier (Chapt. 4.35, 4.36); the top of these two members coincides with the climax of coarse epiclastic deposition in SW Snowdonia. Correlative successions in NE Snowdonia indicate advance of a coarse-grained delta (Bochlwyd Mbr., Appendix sections D1, E9) derived from erosion of the uplifted rhyolitic plutons and volcanic ash about Cwm Dylli (Chapt. 3.4.2).

Directly (40 stratigraphic metres) subjacent to the Pitts Head Tuff, however, the lateral variations in grain size and facies, and the pattern of palaeocurrents indicate development of numerous small alluvial fans

throughout Snowdonia. These comprise the Gwastadnant, Cae'r gors, Y Braich, Cwmystraddlyn, Upper Moel Ddu and Nantmor depositional systems (Fig. 6.11b). Along the NW margin of the Beddgelert depocentre the proportion of debris flow to stream deposits in depositional systems increased to the NE towards Gwastadnant. The disruption of the previous larger-scale depositional systems and reduction in catchment areas along this side of the basin is considered to be a local response of near-surface fractures to the final stages of magma emplacement. The synchronous development of alluvial fans on the opposite margin of the Beddgelert depocentre, with highlands to SE of Nantmor and Yr Arddu fractures, suggests that topographic changes might be a consequence of subsidence of the central basin relative to bounding fault blocks. This volcano-tectonic "trough" largely contained the Pitts Head tuff outflow in SW Snowdonia and resulted in the ramping of the PHT at Moel Ddu against the Nantmor fracture (Enclosure 3).

In NE Snowdonia at Llyn Idwal (App., Section U1,U3; Fig. 6.11b) the tuff is underlain by a coarsening upward succession of fine- to very coarse-grained pebbly sandstone. Sedimentary structures within proximal deposits are dominated by high angle trough and swaley cross-stratification, with planar horizontal lamination in finer lithologies. By analogy with other coastal sequences (Chapt. 5.3), sediments are thought to reflect progradation of a wave and storm dominated shoreface; this would reflect the asymmetric redistribution of material supplied to the Gwastadnant delta front. Sedimentary lithofacies directly beneath the most distal welded outcrops of the tuff at Ogwen Cottage (App. section U3, U4) and Llyn Bochlwyd (section D1, D34) indicate a lower shoreface to transitional offshore setting which supports suggestions from brachiopod palaeocommunities that water depths did not exceed 20-30 metres. As the PHT is *ca.* 30 m thick at this locality, it is clear that water was never

deep enough for the welded tuff to be entirely subaqueous upon emplacement even at its most distal expression.

The transgressive shoreface sequences (Chapt 5.3.4) which overly the PHT reflect gradual retreat of the shoreline toward Snowdon after tuff emplacement. At Cwm Cneifion and Llyn Bochlwyd the sediment above the tuff quickly passes into facies typical of a storm-dominated outer shelf, a deeper water environment than that seen immediately below the tuff (Fig. 6.11 b,c). This change implies rapid sea floor subsidence (*ca.* 80 m) after tuff emplacement. The total amount of subsidence increases to the E. About Bwlch Trimarchog over 1500 metres of marine sediment were deposited between the top of the lower PHT outflow sheet and base of the Snowdon Volcanic Group, compared with 250 m at Llyn Idwal or 2 m of pebbly sandstone about Clogwyn D'ur Arddu (Fig. 5.42).

6.2.5: ASSOCIATED WITH LOWER RHYOLITIC TUFF

The Lower Rhyolitic Tuff Formation (LRTF) represents the lowermost formation of the Snowdon Volcanic Group in central Snowdonia, and contains a minimum of 54 km^3 of acidic tuff. Two phases of eruption can be identified, with the initial volcanic events producing welded sheets(s) of ash flow tuff from sources to the SW Snwdonia (Fig. 6.12a). The formation largely rests on the Pitts Head Tuff (PHT) to SW of Snowdon. The absence of any intercalated sediments from Moel Hebog to Cwm Llan, a distance of 10 km, suggests that this area was subaerial. Where a basin did exist to the NE, over 200 m of sediment was deposited between the PHT and LRTF as noted above. As was previously recognised (Howells *et al*, 1986, p.412), the fining upward sequences directly beneath the Snowdon Volcanic Group at Pont-y-Gromlech, Cwm Cneifion and Pen Llithrig-yr Wrach imply a marine transgression prior to emplacement of LRTF eruptions at these localities. At Pont-y-Gromlech (near the caldera margin), the sequence underlying the LRTF is interpreted as representing a transgressive, non-barred, wave

influenced shoreline (Fig. 6.12b; Orton, 1987; enclosure 2). Facies assemblage immediately beneath the LRTF is typical of the lower shoreface-to-offshore transition zone (water depths of 9-18 m off California, Howard & Reneick, 1981) or inner shelf (Bourgeois, 1980). It is suggested that the maintenance of an uplifted 'plateau' to the SW and matching subsidence to the NE were both related to the lateral migration and accumulation of the buoyant LRTF magma body, with the Llanberis pass and Idwal fault zones of structural weakness accommodating differential movement between fault blocks. The caldera only foundered during its second stage of evolution with development of the assymmetric downsag volcanotectonic trough (Howells *et al*, 1986).

6.3: THE VOLCANO-TECTONIC CONTROL ON SEDIMENTATION

The above consideration of changes in basin configuration and depositional environments in relation to the major pyroclastic eruptions illustrates the strong control of volcanotectonic uplift and subsidence on basin evolution (Fig. 6.13). On a regional scale, before major eruptions, the emplacement and vesiculation of the rhyolitic magma bodies in mid-crustal levels resulted in the tilting of fault blocks, and as a result palaeoslope, away from the sites of magma introduction. The more local changes in basin palaeogeography (*eg.* reduction in sediment supply to delta, more intensive intrabasinal faulting) that <u>always</u> occurred at a stratigraphic distance of 20-30 metres (*ie. ca.* 50,000 years) beneath tuff horizons are thought to reflect the penultimate stages of magma emplacement into a fractured continental crust. There is a direct correlation between the amount of pre-eruption uplift and the size of individual volcanic events. The Pitts Head Tuff and Garth tuff show the most uplift and the best agreement between palaeogeography and independently inferred (*ie.* thicknesses of ash flow tuffs, pyroclastic flows facies etc.) sources of the ash flow tuffs.

The main pyroclastic eruptions and partial evacuation of magma chambers resulted in basin subsidence on both a local (*eg.* the Pitts Head Tuff volcanotectonic trough, the Lower Rhyolitic Tuff caldera) and regional scale (above the Gwern Gof Tuff, Garth Tuff, the lower outflow sheet of the Pitts Head Tuff). The amount and rate of post-eruption subsidence is again proportional to the magnitude of the preceeding volcanic eruption. In addition, regional variations in the amount of subsidence above the Garth and Pitts Head Tuffs, with more post-eruption detumescence further from source, suggest migration (a vacuum effect) towards the eruption sites at Llwyd Mawr and along Conwy fault respectively. In a similar tectonic situation in the Rio Grande rift, Reilinger & York (1979) have suggested subsidence which is similar in size but opposite in sign to uplift in the Socorro area of New Mexico, may also be related to migration of magma; this movement of shallow crustal magma bodies resulted in a progressive increase in the gradient (1m/km to 7.1 m/km) of rift drainage systems from late Pliocene to Holocene times (Bachman & Mehnert, 1978). The Caradocian magma bodies in North Wales were likely similar elongate beams or dikes controlled by the dominant structural grain (*cf.* Cordell *et al*, 1985).

The different styles of volcanic eruption (*ie.* volume, frequency) at Yellowstone and the Taupo volcanic zone have been recently ascribed primarily to the nature of the continental crust (Wilson *et al*, 1984). On a smaller scale, several authors (*eg.* Swanson & McDowell, 1985) have suggested that the structural integrity of the downdropped caldera blocks controls the style of resurgence; if the caldera undergoes multiple collapse episodes, intracaldera basement will be fractured so extensively that it will not contain magmatic pressure during resurgence. Combining these two ideas, it is suggested that the variance throughout the evolution of the Welsh basin in the sizes of ash flow tuffs and corresponding amounts of uplift can <u>BOTH</u> be related to the control of

intrabasinal fracturing on continental strength and flexural rigidity. Strong continental crust would permit the formation of high-level magma chambers; eruptions would be of a large size with substantial pre-eruption uplift (*eg.* Garth tuff, Pitts Head Tuff). Significant intrabasinal fracturing stemming from the successive volcanotectonic changes, on the other hand, would prevent formation of large, shallow magma chambers and eruption size and amount of pre-eruption uplift would progressively become smaller (*eg.* Racks tuff, 3rd and 4th volcanic members of CCVF). The pattern of uplift and subdidence associated with the Lower Rhyolitic Tuff Fm. (LRTF) can be attributed to both extensive fracturing of the upper crust from the numerous "ups and downs" associated with previous volcanic events, and a tectonic stress regime which permitted substantial lateral migration of magma.

In the geological record, the amount of pre-eruption uplift can similarly be related to the strength of the continental crust (Fig. 6.14). In basin with thick or old continental crust, uplift will be volumetrically most significant. It would be spread, however, over an extremely large area making its detection from changes in structural geometries and/or variation in depositional systems almost impossible. At the other extreme, the structural weakness of young continental crust would prevent formation of high-level magma bodies. It is suggested that volcanotectonic uplift and subsidence will be most visible in areas of intermediate crustal age and strength. Here, high level magma bodies can form and pre-existing zones of structural weakness will facilitate movement of crustal blocks; isostatic response of lithosphere to emplacement and evacuation of magma chambers will be compensated more locally, on a scale more suitable for geological observation.

Chapter 7

PRIMARY & SECONDARY PYROCLASTICS: A PRE-DETERMINED DESTINY

> "Each of them escaped from the relationship in which I tried to enclose it, isolated itself, overflowed. I was aware of the arbitrary nature of these relationships which I insisted on maintaining in order to delay the collapse of the human world of measures, of quantities, of bearings; they no longer had any grip on things."
>
> *Sartre, 1938, Nausea*

7.1 INTRODUCTION

Primary and secondary pyroclastic rocks in North Wales include everything from basaltic tuff cones to large, widespread ash flow sheets. It is obviously beyond the scope of this thesis to provide a comprehensive analysis of the deposition of each of these pyroclastic units. As large ash flow sheets have been cited (Fisher & Schmincke, 1984, p. 293) as the best documented examples of subaqueous welded pyroclastic flows, it is pertinent to reconsider this hypothesis in the context of the new interpretations of depositional environments and basin bathymetry (Chapt. 4-6), and identification of more distal correlatives of several pyroclastic units (Chapt 2.2.4.4). A secondary objective of this chapter is to show how the crustal strength argument introduced previously can also be used to explain the style of volcanic eruption. To avoid reiteration of the voluminous outpourings of existing literature on volcanic units (*eg.* Howells *et al*, 1985; Reedman et al, 1987 {Enclosure 3}), mesoscopic and microscopic features within pyroclastics will be only examined in proximity to identified coastlines.

7.2 PYROCLASTIC UNITS

Ash flow tuffs will be described in relation to their depositional context in ascending stratigraphic order (see Fig. 6.3).

7.2.1 GWERN GOF TUFF

The Gwern Gof Tuff comprises a distinct 40 m thick rhyolitic deposit in the Tryfan anticline which quickly pinches out to a thin siliceous siltstone horizon to the north at Craig yr Ysfa (section F1, 10-12 m) and Carnedd Dafydd (section F7, *ca.* 50 m). It overlies shoreface sequences in the Tryfan anticline (Chapt. 6.2.1) and tidal mudflats at Craig yr ysfa (Appendix, section F1), with deposition in more offshore environments to the west. The emplacement of the ash flow induced formation of large sandstone load balls (Fig. 7.3) of sandstone within subjacent strata; related liquefaction has destroyed all sedimentary structures within 3 metres of the base of the tuff. The common infill of pumice and small vugs within the tuff by vapour phase crystallization of REE phosphates and silicates (Fig. 7.2), and evidence for welding (*eg.* Bwlch Cyffryn Drum, Fig. 7.1; M. Howells, pers. comm., 1986) imply that the ash flow was locally hot on emplacement. Although the thinning of the ash flow towards Craig yr Ysfa can be explained by a rhyolite dome at this locality which prevented basin subsidence, the rapid westwards transformation of the ash flow from a 40 m thick sand-grade deposit to a diffuse series of tuffaceous siltstones is attributed to explosive interaction between the hot pyroclastic flow and seawater.

7.2.2 GARTH TUFF

The Garth tuff extends as a continuous primary ash flow sheet for a distance of some 20 km south of its proposed vent on the Conwy fault (Chapt. 6.2.2) to Dolwyddelan whereafter it occurs as isolated pods of tuff (Fig. 7.5, 7.6). If interpretations of depositional environments are correct: 1) the ash flow sheet only travelled at maximum of 3 km <u>into</u> water rather than 15 km as supposed by Howells *et al* (1985) and 2) it never travelled as a continuous sheet <u>underwater</u>. The occurrence of storm-produced hummocky-cross stratification associated with all distal pod facies suggests maximum water depths of 50 m.

The formation of the irregular "transgressive" base to the ash flow

tuff has been the topic of much discussion (Francis & Howells, 1973; Kokelaar, 1979). This feature first becomes well developed along Gallt yr Ogof ridge where regularly spaced (100-150 m) lobate protrusions of tuff extend distances up to 40 m into subjacent braidplain sediments (Fig. 7.4, 7.7), with the asymmetry of the intervening sandstone flames (Howells & Leveridge, 1980) indicating a component of flow to the west at this locality. In the Capel Curig anticline (Fig. 7.8), similar lobate protrusions occur which are larger in size (up to 80 m) and of wider spacing (500-800 m) transverse to probable flow directions. Large "load-balls" of welded tuff which are completely detached from the parent ash flow sheet comprise another manifestation of tuff deformation in the Capel Curig anticline; well developed columnar jointing (Fig. 7.10) indicates these structures were still hot during their emplacement. Analagous loading and detached tuff ball features can be recognized near Dolwyddelan. Liquefaction has disturbed sediments for significant distances (minimum of 6-10 m) beneath the parent ash flow tuff in all these localities.

Further to the south, the isolated pod facies represent the most obvious (see below) component of the Garth tuff. The pod facies have been adequately described by Howells *et al* (1985). It is sufficient to note here that: 1) pods vary in thickness from 2 to 30 m, 2) they decrease in size towards the south, and 3) all tuff pods described are welded.

In the Capel Curig anticline, the author concurs with Francis & Howells (1973) that vitroclastic textures vary within the flow from parataxitic welding near the base (Fig. 7.11) to eutaxitic fabric above, with the top portion largely unwelded (see Chapt. 5.3.2). Perlitic fractures are commonly developed throughout (Fig. 7.11, 7.12) attesting to the rapid cooling of the flow; siliceous nodules which represent both infill of vesicles and later concretions (Fig. 7.13) increase in abundance and size upwards; they are attributed to the upwards migration of silica within ash

flow tuffs shortly after their deposition (Howells & Leveridge, 1980).

7.2.3 RACKS TUFF

The Racks tuff first encountered a wet substrate at Craig yr Ysfa where it flowed into the muddy lagoon (1-3 m deep) envisaged to have occupied most of the Tryfan fault block (Fig. 6.9a, 7.15). The thick (*ca.* 40 m) pyroclastic breccia and bedded tuff that developed at its base upon entering water (Howells & Leveridge, 1980) would probably represent quenched material formed in the early stages of the flow at the shoreline. In the Capel Curig anticline, the tuff varies from 10-30 metres in thickness (Francis & Howells, 1973); in places the underlying mudstones and subjacent sandstones cut almost entirely through the ash flow tuff (Fig. 7.15, 7.16), or the tuff becomes entirely discontinuous (large A in Fig. 7.9). At its basal contact, a delicate and diffuse intercalation of tuff and subjacent sediment is discernible (Fig. 7.14). The tuff displays considerable spatial variation in the degree and amount of welding developed (Francis & Howells, 1973), with both parataxitic and eutaxitic welding textures visible; the former is again the prevalent fabric near the base of the ash flow tuff.

7.2.4 THIRD and FOURTH MEMBERS

The third and fourth volcanic members of the Capel Curig Volcanic Fm. are extremely restricted in their occurrence (Fig. 7.19) with depositional volumes (*ca.* 2-5 km^3) being significantly less than either the Garth or Racks tuff. They stratigraphically interfinger with each other in a complex fashion, particularly about Ffynnon Lloer (SH_660_620) with depositional relationships indicating coeval eruption from several vents. Although they are geochemically similar with considerable overlap in trace element compositions (Fig. 2.30), they differ in both depositional facies and style of inferred eruption. The third member comprises an accumulation of thin primary, often welded, ash flow tuffs (Fig. 7.20, 7.21) which vary in thickness from 4-8 m and appear to have erupted from a single vent about

Carneded Llewelyn (Howells & Leveridge, 1980).

In contrast, the 4th member is dominated by secondary, reworked pyroclastics, although a few thin (less than 8 m) partially welded, vitric-rich ash flow sheets are also found. Over 800 metres of detailed vertical section were logged through the 4th member, some of which is included in the appendix (sections F9, U5). A detailed analysis of sediments and correlation of depositional events will eventually be prepared for publication elsewhere. It will suffice here to note that:

1) "Background" facies indicate that the dominant environment changed from subaerial about Fynnon Lloer in the NW, to shallow marine in the Capel Curig anticline and further to the SE where shoaling ?storm waves (Fig. 7.32) redistributed sediment between volcanic eruptions.

2) Pyroclastic facies include surging high-density sediment gravity flows (Fig. 7.22, 7.26), pumiceous debris flows (Fig. 7.24, 7.25), thin sheetflows (Fig. 7.27), low angle cross-stratification (Fig. 7.28) and airfall tuff. It is difficult to ascertain whether these facies are subaerial or subaqueous.

3) Depositional facies are asymmetrically distributed about the dominant vent (Howells & Leveridge, 1980) at Glyder Fach. To the south and east, coarse-grained crystal and lithic rich pyroclastic surges and debris flows dominate (Fig. 7.22-7.25), with evidence of airfall activity restricted to a few normally-graded turbidity currents containing accretionary lapilli (Fig. 7.29, 7.30). Facies comparable distances to the north of the vent are finer-grained (Fig. 7.26), dominated by vitric-components, commonly pumiceous, and contain accretionary lapilli throughout a thick (50 m) sedimentary succession (Fig. 7.31).

This asymmetric distribution of pyroclastic flow and fall facies on either side of a volcanic event is not uncommon; a direct analogy can be drawn with the Lesser Antilles Arc (Sigurdsson *et al*, 1980), where the

prevailing wind pattern and depositional slope control processes of sedimentation. In North Wales, the greater abundance of fine-grained airfall facies to the north of Glyder Fach provides further supporting evidence for the earlier deduction from shallow marine environments (Chapt. 5.5.2) that the strongest winds in the Ordovician came from the south. As gradients on the 4th member volcanic cone were probably greater towards the marine basin near Capel Curig (SE), than in the direction of the uplifted, subaerial 3rd member volcanic cone to the NNE at Carnedd Llewelyn (Fig.7.19), it is not surprising that volcaniclastic debris flows preferentially developed on slopes facing south from the Glyder Fach vent.

7.2.5 PITTS HEAD TUFF FM.: THE LOWER OUTFLOW TUFF

The lower outflow of the Pitts Head Tuff (PHT) extends as a sheet from Moel Hebog (SH 565 470) to Bwlch Craig (SH 697 629), a distance of *ca.* 25 km from its proposed vent at Llwyd Mawr (SH 508 460). Analysis of depositional sequences within subjacent strata (Chapt. 6.2.4; Fig. 6.9b), indicates that the tuff entered the sea about Bryn Du in LLanberis pass (SH_635_570) and Llyn Clyd (SH_635_597) and continued to flow an additional 3 and 7 kilometres (uncorrected for tectonic shortening) respectively past the coastline at these two localities (Fig. 7.33).

Along Llanberis pass the PHT rapidly changes from a 50 m thick welded deposit at Gwastadnant and Bryn Du (Fig. 7.36) to a series of unwelded sediment gravity flows a mere 2 kilometres further to the SE at Cwmffynnon. Depositional facies within subjacent strata (Gwastadnant Mbr., Chapt. 4.3.7.3) indicate a transformation from alluvial fan through storm-dominated shoreface to offshore environments along this transect. Near Llyn Cwm-y-ffynnon the shoreface successions were locally rotated by listric faults synchronous with tuff emplacement. The deformation was probably induced by loading of the PHT onto an unstable substrate; there is no sedimentary indications of any such topographic changes prior to the

volcanic eruption. Tectonic movements are not envisaged, however, to have resulted in substantial accentuation of depositional slopes or deepening of environments; depositional facies (Chapt. 4.3.7.3) imply maximum water depths of *ca.* 40 m for the most distal PHT successions near Cwmffynnon (Fig. 7.36).

At this locality (F in Fig. &.36), the PHT comprises 55 m of fine- to very coarse-grained rhyolitic volcanic ash. Contrary to impressions given on the recent 1:25,000 map of the area (Howells *et al*, 1985) there are no obvious intercalations of non-volcanic siltstone contained within the succession, and the total thickness of the volcanic component present is comparable to that of the PHT elsewhere in Llanberis pass. Deposits from both debris flows and high-density pyroclastic surges can be recognized. The former lithofacies dominates, with depositional units varying in thickness from 50-300 cm (160 cm average). The individual units display double grading (inverse and normal grading of pumice and lithics respectively) and occasionally display low angle undulating cross-stratification in their basal portions. Although there is little overall pattern to the thickness of these ash beds, examination of internal facies reveals an interesting textural trend which mimics an ideal depositional sequence (Fig. 7.64,7.65) that one would expect from emplacement of an intermediate volume pyroclastic flow. For example, the lowermost debris flows have unusually high crystal contents (*ca.* 20%; Fig.7.43, 7.44) for the PHT (*ca.* 5% is maximum usually observed) and frequently contain extremely large (for the PHT) angular lithic blocks of fine-grained rhyolitic ?tuff or even fine sandstone (up to 50 x 100 cm size observed). In contrast, the uppermost debris flows have a higher proportion of pumice with few accidental lithic fragments visible. Most pumice shows no evidence of post-depositional flattening and weather to produce rounded, pithy, vugs within the debris flows. Those in the top quarter of the succession, however,

display a distinct eutaxitic fabric (Fig. 7.40-7.42). In the uppermost debris flows this flattened pumice is contained within large (10-40 cm) sub-rounded lithic blocks of partially welded PHT set in a matrix of unwelded volcanic ash (Fig. 7.39-7.41). Previous investigators of this locality have regarded the succession described above as "thin beds of reworked tuff interbedded with [epiclastic] sediments" (Reedman *et al*, 1987). Consideration of textural relationships within more proximal and adjacent portions of the ash flow tuff, however, suggest otherwise.

Near Bryn Du (SH_6405_5674) for example, the bottom 10-15 metres of the PHT is unwelded, and comprises an disorganised collage of amoeboid-shaped "blobs" of various admixtures of fine-grained tuffaceous siltstone, massive unwelded tuff, and very coarse grained sandstones. Large (up to 30x70 cm) lithic blocks of fine-grained rhyolitic ?tuff are common (Fig. 7.38). An even more confusing melange occurs near Gwastadnant (SH_6131_5800; Fig. 7.36). The top portion of the PHT at these two localities comprises a pumiceous, vitric-rich deposit. In adjacent exposures of PHT at Pont-y-Gromlech (SH 628 564), Wright and Coward (1977) have ascribed ductile deformation of welding fabrics and formation of diapir-like structures to upward streaming of water vapour into the flow. The distribution and abundance of siliceous nodules in these nearshore environments contrasts with the situation in subaerial settings where nodules clearly post-date all welding; enormous "cannon-ball" sized concretions (Fig. 7.55) resulted from flux of water along the impervious base of the parataxitically-welded PHT at Moel Hebog and Moel Ddu.

When one pulls together all the various strands of evidence which indicate incorporation of substantial quantities of sediment and water into the base of the Pitts Head Tuff from Gwastadnant to upper slopes of Bryn Du (A to C in Fig. 7.36) one reaches the inescapable conclusion that the rapid transformation in 1 kilometre from a comparatively "normal" ash flow deposit

(at D) to a series of debris flows at Cwmffynnon (F), was a natural consequence of the continuation of this process into deeper water. The alternative is even more inconceivable. The ash flow would have had to "freeze in its tracks" on the "sight" of water, weld together, and later miraculously erode to yield a series of debris flows of comparable total thickness without inverting the stratigraphy (*ie.* pumice/lithic distributions) characterising the parent flow.

Further to the NE, depositional features indicate a contrasting, less destructive pattern of interaction between the pyroclastic flow, subjacent sediment and surrounding seawater during emplacement. East of Gribin ridge near Llyn Bochlwyd, for example, up to 30 metres of vertical movement on a listric growth fault caused substantial thickening of the tuff within a half-graben (Fig. 7.45, 7.46). The only evidence of any change in flow behaviour associated with this tectonic activity is provided by development of a thin (1.5 m) horizon of micaceous and feldspathic siltstone at the base of the PHT (Fig. 7.49) attributed to mixing of the ash flow with any attainable sediment. Subsequent rotational movement along the sole fault produced tensional fissures in the base of the PHT which were filled by several sandstone dykes, with the larger 10 metre wide intrusion recognised as such by F. Fitch during earlier mapping and included on the 1:25,000 Special Sheet of Snowdonia. Both dykes are texturally graded with coarsest detritus (rhyolite cobbles up to 6 cm length) occurring along their centres; a stratigraphic organization is also visible with the fine-grained margins representing resedimentation of the micaceous base to the PHT being later intruded by sand. Whereas the small dyke is internally structureless (Fig. 7.49), the large intrusion commonly displays both a distinct wall-parallel layering from shear during deposition (Fig. 7.50), and transverse bedding in axial regions suggests plug flow (Fig. 7.51).

The lower margins of both sandstone dykes are sharp with the only

evidence within the ash flow tuff of a nearby intrusion provided by drag folding and rotation of flattened pumice clasts. This contrasts with the diffuse compositional gradients near the ends of the sandstone dykes where complete mixing between tuff and epiclastic sediment makes recognition of margins impossible. Numerous apophyses and pods of sandstone (Fig. 7.52, 7.53) adjacent to the dyke in intervening portions record the first evidence of mixing between the tuff and its incorporated sediment. Radial fractures within these sandstone bodies are attributed to rapid contraction following emplacement into a hot ash flow due to loss of pore water as steam; similar textures have been observed within small mudstone fragments incorporated within basaltic lithic fragments of Quaternary phreatomagmatic tuff cones in Germany. The above textural relations indicate that dykes were emplaced after welding had commenced at the base of the tuff but before the top portions were lithified and perhaps were even still being deposited.

This mode of incorporation of both sediment and water during tuff emplacement is not restricted to just this locality. To the west of Llyn Idwal, rafts of sandstone up to 4 m thick (?detached dykes) occur toward the top of the ash flow deposit (Fig. 7.54) where the PHT thickens across a contemporaneous fault. Distribution of siliceous nodules and tuff fabric near Ogwen cottage has also been attributed to upward movement of water vapour through the flow (Wright & Coward, 1977). Although these mixing processes were not explosive, they eventually led to the demise of this lobe of the ash flow. At Bwlch Craig (SH 697 629), the sole representative of the formation, as confirmed by whole-rock geochemistry (Chapt. 2.2.4.4) is a 13 m thick sediment gravity flow, which grades upward from a crystal-rich base to a fine-grained vitric-enriched top. Shard textures are extremely well defined with <u>no</u> evidence of welding visible (Fig. 7.57, 7.58); perlitic fracturing of glass in the centre of the deposit indicates some portions of the flow were locally hot (but see Fisher & Schmincke, 1984, p. 327) on

emplacement.

7.2.6 THE "TUFFITE" "CLAN".

These primary pyroclastic units vary from 7 to 15 m in thickness; to less than 0.18 to 0.375 km^3 (.09 to 0.19 km^3 dense rock equivalents assuming 50% porosity; see Crisp, 1984). For reference, the catastrophic May 18, 1980 eruption of Mt. St. Helens produced 0.12 km^3 of volcanic ash. "Tuffites" consist of cuspate rhyolitic glass shards which vary from medium silt to fine mud in size (Fig. 7.58), and rarer albite-oligoclase or quartz phenocrysts, set in an extremely fine-grained recrystallized matrix. Whole rock geochemistry yields Sio_2 concentrations of 75 to 85%. Sediments are usually organised into 10-60 cm thick, normally-graded beds. Planar to undulating horizontal lamination, low angle cross-stratification (similar to Fig. 5.16), and "slurried" soft-sediment deformation structures (Fig. 7.59) are common internal structures within proximal settings and attest to their rapid flow velocities and sedimentation rates. The mode of sediment transport would be analogous to a high density turbidity current, with occasional internal breaking waves related to supercritical to subcritical transformations in flow regime. In a most cases, the depositional context suggests entry of dilute pyroclastic flows into the sea, or collapse eruption columns into 10-20 metres of water. The lack of large scale organization of bed thickness or grain size for most pyroclastic units implies little variation in eruption dynamics. In only one case (Tuffite Mbr., Tryfan anticline, Fig.5.61, 7.60, 7.61) can lithofacies (*ie.* each bed and entire successions normally-graded) be used to suggest a waning eruption column (*cf.* Fiske & Matsuda, 1964; Yamada, 1973).

The "tuffites" often form the initial phase of bipartite pyroclastic eruptions, being overlain at Cwm Cneifion (Fig. 5.44) and Llyn Cowlyd (Fig. 5.56) by coarse sand-grade volcanic ash. This eruption pattern mimics on a small scale the gross depositional succession resulting from the overlapping

eruptions of the 3rd and 4th members of the Capel Curig Volcanic Fm..

7.2.7 LOWER RHYOLITIC TUFF

Although the LRTF was not examined in any detail, a few relevant depositional features are worth noting (see Fig. 6.12). These are:

1) The pyroclastic flows from the earliest eruption(s) in the SW disintegrated upon entry into the sea about Snowdon (see Orton, 1987; they perhaps generated the series of unwelded rhyolitic siltstones near the top of the depositional succession in Cwm Cneifion (Fig. 5.33).

2) The main eruption from volcanic vents along Llanberis pass (Rast *et al*, 1958; Howells *et al*, 1986) was immediately preceded by a phase of NW-SE extension which erupted local accumulations of pillow lavas and hyaloclastites from a vent along the nearby Glyder Fawr fracture.

3) The true water depth of the shallow marine caldera can only be determined between the debris flow-producing volcanic episodes when reduced rates of sediment supply allow the "background" marine processes to express themselves in the depositional record. During one such hiatus, the development of high angle trough cross-stratification containing *Dinorthis berwynensis?* brachiopods (BGS internal report 80/24 #3) suggest a shallow-water (10-30 m) high energy environment.

4) Comparison of brachiopods within the LRTF between two limbs of the Idwal syncline, with persistence of *Dinorthis* in the west and sowerbyellids in the east (S.P.Tunnicliffe, BGS internal report PD/80/27) indicates continuation of the earlier pattern (Chapt. 5.4.2) of greater subsidence to SE of Idwal fault.

7.3 ENTRY OF ASH FLOW TUFFS INTO THE SEA

7.3.1 SUBAERIAL PYROCLASTIC FLOWS

Subaerial pyroclastics have traditionally been divided into a tripartite genetic classification of (1) fall, (2) flow and (3) surge (Sparks & Walker, 1973). Much discussion has recently been generated over

the classification of individual pyroclastic units, with the May 18, 1980 Mt. St. Helens deposit being a classic modern example. This has been attributed to emplacement by a pyroclastic surge (Moore and Sisson, 1981), a complex combination of flow and surge (Hoblitt *et al*, 1981), a pyroclastic flow (Walker, 1983) or most recently a blast surge (Fisher *et al*, 1987). As a extension of ideas developed regarding the spectrum of flow behaviours exhibited within subaerial alluvial deposits (Chapt. 4.4.2.1), it seems more logical to recognise that the three classes of deposit represent end-members of a triangular continuum of processes (Fig. 7.62). Pyroclastics "flows" would mostly inhabit the flow-surge join, with positional variation in flow mechanics and resultant facies attributed to the amount of fluidization (see Wilson, 1985). At one end of the spectrum, small volume, dense, high aspect ratio (*sensu* Walker, 1983) ignimbrites appear to behave as viscous Bingham mudflows to form small structureless lobes deposited *en masse* (Fig. 7.63). These grade transitionally into more dilute, highly inflated, Newtonian, gas-particle mixtures where fluid turbulence is the dominant transport mechanism; the resulting bed-by-bed deposition from traction and suspension sedimentation produces bedforms (*eg.* Self *et al*, 1984; Wilson, 1985) physically and genetically indistinguishable from structures of density stratified surges (*eg.* Valentine, 1986). Most of the pyroclastic flows described in the literature would represent the middle portion of this spectrum, behaving as semi-fluidized, quasi-Newtonian, fluid with little or no yield strength (Sparks, 1976). A deflated pyroclastic flow moving in a <u>laminar</u> fashion will cohesively freeze <u>from the top down</u> when velocity-induced shear stress drops below the critical yield strength to progressively form a <u>plug</u> near the top of the flow. The "standard ignimbrite flow unit" resulting (Fig. 7.64, 7.65) would comprise a thin inversely graded base (unit 2a) with the main body (unit 2b) displaying reverse grading of pumice and normal grading of dense, lithic clasts in

"textbook" examples.

As with all sediment gravity flows, a pyroclastic flow can be conceptually divided into a head, body and tail (Wilson & Walker, 1982). The head region and base of the body are the critical area of interest which will govern the ability of a flow to successfully enter water. Although they are dynamically coupled, it is convenient to consider them independently for purposes of discussion. The head is the controlling feature of the flow. By analogy with other sediment flows, it will carry an overhanging nose (Fig. 7.66) due to the viscous response of the ambient medium to the intruding fluid (Allen, 1971). Most authors envisage that substantial quantities of air are ingested within the head of the flow (Sparks, 1976; Wilson, 1985) with the amount of air incorporated directed proportional to flow velocity (Wilson & Walker, 1982, p. 589). Air ingestion can occur both beneath the overhanging flow front and as plume-like vortices of gas/particle mixtures in the turbulent wake (Fig. 7.67) of the flow head (Denlinger, 1987). Experimental evidence has shown that slopes of at least 0.5° are necessary for gravitational forces along the slope to balance friction and entrainment drag to produce a steady flow front velocity (Britter & Linden, 1980). The rate of mixing at the flow front, and growth in head size increases dramatically with slope (Fig. 7.68). At a slope of 20°, for example, direct entrainment from the surrounding medium accounts for one-third of the growth of the head with the rest added from the current behind. In a laboratory simulation of the movement of pyroclastic flows down slopes, Huppert *et al* (1986) also observed that the front of a flow was "quite unsteady" and dense bulbous clumps of relatively undiluted fluid (which eventually mixed with ambient fluid) moved out ahead of the main flow. These pulses were an intrinsic result of non-linear density mixing at the flow front which periodically provided protection from direct mixing with the environment (compare with snout of debris flow shown in Hampton,

1972, fig. 9).

The second "problem area" is the basal boundary layer of the flow (Fig. 7.69). Here, the forward motion of the flow is affected by the frictional resistance exerted by the underlying sediments. The flow mechanics in this region will determine the manner in which steam generated within substrate mixes with the pyroclastic flow. If there are no large turbulent eddies in the free stream (*ie.* the pyroclastic flow travels in laminar fashion) than a laminar boundary layer might persist for "a long time" (Middleton & Southard, 1984, p.3-20) It seems intuitively obvious, however, that through trapping of eddying motions inherited from formation of the pyroclastic flow, and continual accentuation of these disturbances by roughness in the substrate that the boundary layer will be turbulent. Similar conclusions were recently reached by Valentine & Fisher (1986) who demonstrated theoretically that velocities necessary to generate fluid turbulence within the base of a pyroclastic flow are commonly exceeded; they thought that the layer 1 deposits of ignimbrites (Fig. 7.64) were the direct result of this flow regime. The base of a pyroclastic flow was envisaged as a place where body transformations (*sensu* Fisher, 1983) were continually taking place. At Mt St. Helens, Kiffer & Sturtevant (cited in Valentine, 1987, p.620) have reported evidence based on erosional vortices that the turbulent boundary layer in some areas was on the order of 3-10 metres thick. Although the high particle concentrations in subaerial pyroclastic flows would dampen any significant upward thickening of this turbulent "carpet" to involve the entire flow, this will not be a restricting condition in subaqueous environments (see below).

There is a deterministic order to the chaos generated by the turbulence at the base and front of a sediment gravity flow (Allen, 1985, p.106) and full comprehension of the mixing of ash flow tuffs upon entry into the sea must consider flow dynamics in 3-dimensions. The viscous sublayer is a thin

(mm size) layer of flow next to the boundary (Fig. 7.69a,b) where viscous shear stress predominates over turbulence. If the substrate is dynamically "rough" than no sublayer will be present. Even if a sublayer exists, flow is not entirely laminar, however, but organised into a series of contra-rotating streamwise vortices (Fig. 7.69c) which periodically "burst" into the overlying outer flow. Similar flow structures, which are probably genetically related (Allen, 1985, p.121) give rise to corkscrew vortices which can effect the entire flow; Langmuir circulation cells which result from complex frictional interaction between wave and wind-related currents (Fig. 7.70) are an excellent example. On a even larger scale, thermal instability in atmospheric boundary layers has been used as an explanation for the oppositely spiralling wind patterns involved in the formation of seif dunes on deserts (Fig. 7.71). The last type of secondary flow is directly comparable to pyroclastic flows where Rayleigh-Taylor gravitational instability and unidirectional shear are also combined. In any sediment gravity flow, these corkscrew vortices will result in a spanwise series of regularly spaced clefts/lobes along the gravity current head (Fig. 7.72). Fisher (1979) postulated that internal vortices are an integral part of an advancing pyroclastic flow and cites observations by Taylor on the eruption of Mt. Lamington that nuée ardentes tend "to expand laterally in vortex like forms"; he recognizes that they will be most important in the dilute top of the pyroclastic flow. It is proposed here, however, that the lobe and cleft configuration will continue for considerable distances into the main body of the flow behind its advancing front, and provide a viable mechanism for incorporation of air and/or water throughout the entire flow. The extent to which lobe and cleft structure develops will depend on flow thickness and velocity (Fig. 7.73). Shear stress exerted by a flow will vary transversely along the flow front according to the different mean velocities associated with these lobes and clefts; Allen (1971) has related the periodic spacing

of flute marks at the base of turbidity currents to the greater erosive power beneath lobes. In a volcanic example, Fisher (1971) attributed regularly spaced (5-25 m), 3 m deep radial grooves and associated ridges on the flanks of Koto crater, Hawaii to secondary flow along the polylobate front of a surge.

The above "snapshot" of the front and base of a pyroclastic flow has only considered spatial variation. All flows naturally evolve, however, as they travel away from their source (Fig. 7.62). Transformation in flow behaviour is related largely to internal processes such as gravity segregation of a basal avalanche from a collapsing eruption column or elutriation of fine ash from a flow. Turbulence is most likely in the early stages of the flow when the gas-particle mixture is highly inflated and travelling fast (Sparks, 1976, p. 177). As the flow deflates further from source, the increasing concentration of particles in the basal avalanche suppresses movements under gravity or fluid forces; with increasing viscosity and flow strength, and decreasing velocity, the semi-fluidized pyroclastic flow will eventually develop. In the Taupo ignimbrite, fluid turbulence dominated flow mechanics for 13 km from the vent (Wilson, 1985); with smaller volume (*ca.* 0.6 km^3), less energetic, pyroclastic flows associated with Maar volcanoes in Germany the transition from turbulent to laminar motion occurred at a distance of 2 km (Freundt & Schmincke, 1986).

The order to this progressive evolution, however, is compounded in the real world by external stimuli such as changes in topography. In Germany, for example, small pyroclastic flows associated with a maar volcano repetitively experienced hydraulic jumps (supercritical-subcritical flow transformations) related to changes gradient or flow deflection during their travels (see Fig. 7.74). Although it is uncertain whether the density stratification and yield strength of the basal avalanche permitted the same increases in turbulence and related flow thickness (roughly 2.4 times;

Komar, 1971) found in marine settings, the significant reductions in velocity (*ca.* 30%) and flow competence associated with the flow transformation resulted in rapid depositional separation of discrete lenses of large lithic blocks (Freundt & Schmincke, 1985; Fig. 7.75).

7.3.2 DISINTEGRATION IN NORTH WALES

Ash flows in North Wales, which all must have had densities greater than 1 kg/m^3, all disintegrated upon entry in to the sea to either generate fine-grained phreatomagmatic ash, or a series of unwelded subaqueous debris flows. When one considers a pyroclastic flow as a semi-fluidised gravity current with a turbulent and erosive head, and turbulent basal and upper boundary layers as reviewed above, it is not surprising that the Welsh ash flows experienced a few "mid-life crises" upon encountering water. Sedimentary facies and brachiopod palaeocommunities indicate that the maximum depth of water in which welding occurred was 40 m. The similarity of this value to the deflated flow thickness is not though to be a coincidence. The contrasting patterns of flow disruption and style of mixing with seawater can be related to inferred flow temperatures, flow velocities, depositional slopes, amounts of fluidization and related morphology of the flow front. Discussion will focus on the Garth and Pitts Head Tuffs.

As these ash flows were respectively 17 km and 10-13 km from the inferred vents when they reached the coastline, and obviously less energetic than the Taupo eruption being confined by topographic reliefs of 200 or 300 metres (Chapt. 6.2.2), it was probably safe to assume that they were largely deflated during this stage of their evolution. The high grade (welding) of both tuffs, absence of any surge type bedforms (but see Fig. 7.50) in subaerial deposits, absence of a significant precursor plinian phase, and evidence for primary welding (see below) suggests that "deflated" flow mechanics were largely inherited from low, fountaining, eruption columns (if

at all) which ingested only limited amounts of indrawn air during column collapse (Sparks & Wilson, 1976).

In a subaqueous setting, two additional parameters which might modify the flow dynamics and mixing behaviour must be considered. The first of these, the Leidenfrost effect (Fig. 7.76), has long been a favourable mechanism in attempts to use the Garth tuff as a model for non-disruptive entry of ash flow tuffs into the sea (*eg.* Howells *et al*, 1985). Although film boiling (Mills, 1984) might assist the quiet non-explosive and non-clast forming incandescent effusion of viscous lava under water, and perhaps even aided separation of the Capel pods (see below) from the main sediment gravity flow, it is rather doubtful whether this sheath can protect a vigorously advancing turbulent flow front from contact with water. It has been recently illustrated (Kokelaar, 1986) that even during the interaction of basaltic lava and water a vigorous mechanical impulse (rapid entry rates or confrontation by a energetic surf zone) would result in violent explosions from steam trapped beneath and within the lava to produce large littoral cones of hyaloclastite, scoria, and lithic fragments. Although steam has a low thermal conductivity (Mills, 1984), it is an excellent convector of heat. The second difference about a subaqueous settings is that once underwater the increasing hydrostatic pressure will reverse the subaerial trend of a progressively deflating pyroclastic flow. Any water mixing at the flow front, rather than being elutriated will be incorporated into the flow to generate both subaqueous debris flows and derivative turbidity currents (*cf.* Hampton, 1972). The flow transformation is directly analogous to the transformation of debris flows to hyperconcentrated flood flows when lahars enter streams (Pierson & Scott, 1985).

7.3.2.1 THE PITTS HEAD TUFF

As noted earlier, the behaviour of the Pitts Head Tuff (PHT) upon entry into the sea varied with distance from its source. It is probably safe to

assume that the velocity of the flow decreased further from source. In conjunction, the flow would progressively become less fluidised and less erosive with the turbulent flow head being of smaller height. Several lines of field evidence support this argument. Near Llyn Ogwen, the frequently occurring sandstone rafts (*eg.* 7.54) are usually supported near the top of the flow; any "lithic clasts" in more proximal settings (*eg.* at Bryn Du, Fig.7.38) occur near the base of the flow. Secondly, whereas at Llyn Ogwen there is little evidence for any erosion along the flow base, the complex textural relations along Llanberis pass suggest the opposite.

An additional difference between the two regions concerns the depositional slope and tectonic activity accompanying flow emplacement. Along Llanberis pass, turbulence is thought to have been enhanced within the flow by: 1) a higher depositional slope (*ca.* 25 m/km, 1.5°), 2) hydrodynamic roughness due to back-tilting of subjacent strata during earliest stages of flow emplacement, and 3) vortices from deflection of flow around subdued remnants of the uplift at Cwm Dyli which had earlier fed the Bochlwyd Mbr. delta (Fig. 6.9b). In addition, a flexural increase in depositional slope across the Beddgelert structural weakness was probable (Fig. 7.78). A change in gradient such as this would cause the pyroclastic flow to hydrodynamically separate from the substrate; the turbulent back eddies produced would be assisted by the large volumes of buoyant steam from trapped water in rapidly transforming the PHT to the unwelded debris flows that are seen. A pulsating flow front, either inherited from eruption dynamics or intrinsically related to non-linear mixing at the front of a density stratified flow, can be used to explain a series of derivative sedimentary gravity flows rather than a single product.

To account for the textural features at the top of the distal succession at Cwm ffynnon one needs to invoke primary welding within PHT. This process involves laminar shear of a compacting and welding mass of

shards to for a primary flow foliation during forward motion of the flow (Chapin & Lowell, 1980). There are numerous textural features which suggest that temperature and flow mechanical conditions were suitable for the development of such fabrics and/or they did form. These include: 1) extremely well-developed parataxitic welding fabric throughout subaerial deposits suggesting accretion in laminar flowage regimes, 2) rotated welded tuff fragments within welded PHT at Moel Ddu (Fig. 7.34), 3) rheomorphic folding of PHT at Pont-y-Gromlech suggests temperature of glassy particles was well above their softening point, and that they locally agglutinated early to form a viscous fluid with a rheology similar to rhyolite lavas. The development of a primary welding fabric was probably inaugurated during laminar shear along the boundary of the volcanic vent(s)(*cf.* Reedman *et al*, 1987) and later enhanced when a concentrated and compressed pyroclastic flow "boiled over" or "topped" the rim of the developing volcano-tectonic trough at Llwyd Mawr (Freundt & Schmincke, 1985; Komar, 1983; Fig. 7.77). It is envisaged that by the time the PHT reached Bryn Du, a small plug of welded tuff had locally formed near the top of the pyroclastic flow. The steam and associated shock waves produced upon entry into the water would immediately fracture the raft into the series of lithic blocks that is seen. It is thought that both abrasion during final stages of transport and volatile streaming after deposition can account for the observed clast rounding.

In contrast to the above, near Llyn Ogwen, any tectonic changes occurred after a considerable portion of the flow had already been deposited. Incorporation of water into an already viscous, ?partially aggutinated, slower moving flow was less disruptive. Continued travel along a comparatively "smooth" and shallower depositional gradient eventually produced the unwelded 13 m thick "megaturbidite" at Bwlch Craig. The thin size of this derivative pyroclastic flow in comparison with the 40-60 m thick parental deposit suggests that even this flow transformation was

destructive. It is proposed that phreatomagmatic explosions from the entry of the Pitts Head Tuff into the sea produced most of the airfall component which overlies the flow at Pont-y-Gromlech and was identified geochemically near Yr Arddu. If this hypothesis is correct, the widely dispersed volcanic ash which would result from such interactions (Self & Sparks, 1978) might provide one of the more useful chronostratigraphic markers in the Caradocian. Unfortunately, Ordovician wind patterns (Chapt. 5.5.3) mean that one has a higher probability of discovering the horizon in the Lake district than the Welsh basin.

7.3.2.2 THE GARTH TUFF

The behaviour of the Garth tuff upon entry into the sea contrasts with patterns of flow transformation experienced by the Pitts Head Tuff. The genesis of the Capel "load-balls" and "pods" is the most crucial component of any model erected to explain the different style of flow disruption. As the Garth tuff was a greater distance (17 km *cf.* 12 km) from its vent upon encountering the stormy Welsh seas, and was associated with a smaller volcanic eruption (28 km^3 *cf.* 54 km^3), it is probably safe to assume a slower velocity and less substantial degree of fluidization at the flow front in comparison with the Pitts Head Tuff. The greater interfacial friction accompanying this scenario would be conducive to the development of a polylobate flow front. As suggested by Branney (1986, his Fig. 1), the development of the lobate protrusions at the base of the Garth tuff would commence immediately upon passage of the pyroclastic flow over a wet substrate, rather than solely by static gravitational instability after flow emplacement (Francis & Howells, 1973). Some field evidence for this belief is provided in the Capel Curig and Tryfan anticlines where: 1) asymmetry of flames between lobate protrusions implies flow movement after their initiation (Fig. 7.7), 2) the parent flow overlying "load-balls" is often not stratigraphically thinned as it should be if the structures were

entirely post-depositional (Fig.7.9), and 3) "pre-natal pods" (ϐ in Fig. 7.9) are often attached by thin southward-protruding "umbilical cords" to the parent flow.

It is proposed that the irregular base to the Garth tuff reflects transverse variation in shear stress along the front of a gravity current associated with lobe and cleft development. If this is true, the spacing of the lobate protrusions would reflect the dynamics of the Garth tuff during transport. At the Reynolds numbers applicable to pyroclastic flows, the limiting conditions of a lobe spacing/nose height ratio of 2.7 (Allen, 1985), and nose-height/head-height ratio of 0.12 (Allen, 1985, p.228; Simpson & Britter, 1979, p. 477, 495) can be used. The lobe spacing of *ca.* 150 m along Gallt-yr-Ogof therefore implies a nose and head height of *ca.* 50 and 463 m respectively. As the head/body ratio in sediment gravity flows is of the order of 0.25 (Simpson & Britter, 1979), one would infer that the Garth tuff had a thickness of 120 m during transport. Its present thickness is of the order of 65-80 metres (Howells & Leveridge, 1980, p. 3 & 10) which suggests the tuff body was 50 to 80% expanded over its final compacted state during transport. This would correspond to flow densities (neglecting pumice and using s.g. = 2.7 g/cm^3 for glass shards) of *ca.* 1.5 to 1.8 g/cm^3. It would be extremely naive to expect the above calculations to be exact. The results are thought to be reasonable, however, as a more inflated flow (limiting value is 170%) would not be contained by the topographic relief of *ca.* 250 m inferred from sedimentology (Chapt. 6.2.2) and also be lighter than water. It is encouraging that comparable values (less than 100%) for the degree of expansion in the distal portions of the Taupo ignimbrite were obtained by Wilson (1985) using a totally different scientific approach based on pumice density data. The larger spacing of the lobate protrusions in Capel Curig anticline can be attributed to exploitation during loading of surface irregularities (Branney, 1986), which in this case would be braid-

bars, floodplains, and abandoned fluvial channels. The density of the sandy substrate (assuming densest packing of spherical grains of quartz to give 26% porosity with no cementation or matrix; Pettijohn *et al*, 1973) was on the order of 2.3 g/cm^3 before tuff emplacement. This is not envisaged to have been a problem, however, as the critical density which the partially inflated flow needs to exceed to allow downward protrusion of the tuff will be that of a fluidised steam/sediment mixture (pers. comm by P. Kokelaar to M. Branney, 1986).

The origin the enigmatic Capel pods can now be examined in context of the above discussion. Two contrasting models have already been proposed relating their development either to 1) "buds" detaching from the flow front at a change in topography (Howells *et al*, 1985) or 2) gravitational instability at the distal end of a high-grade ignimbrite (Branney, 1986). The author concurs with certain elements of both models. Their genesis must be conceptually linked with formation of "load-balls" in Capel Curig anticline (Branney, 1986). On the other hand, the abrupt change from a continuous ash flow sheet to isolated pods certainly implies a change in topography (Howells *et al*, 1986). In addition, the wide separation of distal pod facies (1.5 km) argues against 'drainage' (*cf*. Branney, 1986) from a continuous parent sheet (Howells *et al*, 1986).

Both of the above models tend to treat the Garth tuff as a viscous fluid with rheological properties analogous to a rhyolite lava (flow budding, draining), and assume that the mapped sheet and pod facies are the entire representative of the pyroclastic flow which was emplaced. Their opposing models can be unified if one treats the Garth tuff as a fluidised, vigorously advancing, dynamic sediment gravity flow with a turbulent, polylobate flow front. Upon entry into the sea, the pyroclastic flow would start ingesting water into the base of the flow head, with preferential streaming of volatiles up through the entire flow along clefts. It is

envisaged that this localised mixing pattern was maintained until the nose of the flow started to 'dive' under water. At this point, the additional incorporation of steam and water into the top and wake of the head through Kelvin-Helmoltz instabilities appears to have been 'too much for the flow to handle', and most of the Garth tuff was quickly transformed to a dilute, water-supported, sediment gravity flow.

A basaltic mudflow deposit at Capel Hill (Fig. 7.80), sheds some light on a possible mechanism for the genesis of the isolated pod facies in the context of the above scenario. Although the mudflow horizon is only 5 cm thick, it contains an outsized clast of dense fine-grained rhyolitic tuff which is 25 cm in size. The original flow which transported this clast was obviously at least as thick as the clast itself; comparison within its younger cousin (Fig. 5.55) suggests a flow thickness of ca. 1 metre is more likely. Through a change in flow mechanics, perhaps related to a hydraulic jump and incorporation of water, the sediment gravity flow was unable to transport its isolated dense clasts, deposited them, and continued to flow further (to the N) into the basin. A thin, otherwise insignificant, muddy veneer represents the only marker for the passage of a much thicker volcaniclastic flow. A modern pyroclastic analogy is provided on Augustine volcano, Alaska (Fisher & Schmincke, 1984, their fig. 8-15), where a giant *ca.* 5 metre high block was left stranded on upper volcano slopes whilst the pyroclastic flow in which the block was transported was deposited on the lower flanks of the volcano. With reference to the Capel pods, if the dense "rhyolitic clast" (*ie.* a pod) was hot, it would naturally "burrow" into the subjacent sediment during transport and once deposited. A hot ball-bearing on a cold slab of butter would provide a good analogy. It is therefore envisaged that the Capel pods represent dense "hot spots" (perhaps plugs of primarily welded tuff) from the unmixed centre of the flow lobes. Their greater and increasing density relative to the enveloping

water/sediment mixture, enabled gradual separation from the surrounding pyroclastic flow which continued to travel further into the basin.

If this model is correct, it implies that: 1) further to the south a unwelded fine-grained shard-rich siliceous siltstone represents the dilute part of the Garth tuff which "got away", 2) at an unspecified distance above the stratigraphic level of the pod facies, which would decrease further to the south (*ie.* as the ash flow became colder its pods would sink less upon detachment from the flow), one should expect to find a thin (1-5 m ?) unwelded siliceous siltstone as a marker of the stratigraphic position of the disintegrating ash flow tuff which once carried the pods of hot tuff. Some support for the second expectation of the above model exists within shallow marine successions near Dolwyddelan (Fig. 6.8). Here, a inconspicuous unit of tuffaceous siltstone approximately 50 metres above the stratigraphic level of a Capel pod might represent the unwelded remains of the Garth tuff. Although this field evidence is comforting, it is by no means conclusive; whole rock geochemistry is currently in progress to test the validity of the hypothesis.

7.3.3 SUBAQUEOUS WELDING: SOME LIKE IT HOT

7.3.3.1 WHAT IS IT?

Welding has been defined by Smith(1960) as:

> "that process which promotes the union of glassy fragments" with "the degree of welding [ranging] from **incipient** stages marked by sticking together or cohesion of glass fragments at their **points of contact** and within the softening range of glass, to **complete** welding marked by the cohesion of the **surfaces** of glassy fragments accompanied by their deformation and elimination of pore space"

The distinguishing feature is the plastic deformation of glass shards; the time taken will vary according to sheet thickness (depositional load) and water content, with 0.5% by weight water in glass lowering threshold temperatures by *ca.* 60°C. The degree of welding is also dependent on load pressure with the flattening ratio (F=length/height of pumice) in most deposits steadily increasing from the top downward into ignimbrite sheets

(Peterson, 1979). Theoretical considerations have been presented by Sparks *et al* (1980) which indicated that development of welding within subaqueous pyroclastic flows is extremely favourable. In suggesting that cooling of a subaqueous pyroclastic flow due to ingestion of water (*ie.* 800°C to 755°C) is more than offset by the decreased viscosity (2 x 10^{11} to 2 x 10^{9} poise) of glass as steam becomes dissolved, the authors only considered water which must be drawn into the flow to maintain hydrostatic equilibrium. The experimental evidence (Friedman *et al*, 1963; Shaw, 1963) used by Sparks (op cit) also shows that if more substantial amounts of water are ingested into the pyroclastic flow, and it cools from 800°C to only 500°C during incorporation of water than the viscosity of the hydrated glass will actually increase to *ca.* 10^{15} (Fig. 7.81). This value is in excess of the maximum viscosity at which incipient point contact welding can occur (Rogers & Guest, 1967). As field evidence suggests the Welsh ash flow tuffs did incorporate large quantities of fluid, one must look to another mechanism to account for "welding" textures.

Pressure solution represents a similar process within sedimentary rocks. It occurs when (Pettijohn *et al*, 1973, p. xx):

> "high effective pressures develop at point contacts of quartz grains and increase the solubility at these points such that they preferentially dissolve. The dissolution liberates dissolved SiO_2 to the pore water which then becomes supersaturated and reprecipitates the SiO_2 as quartz overgrowths".

At 25°C, the solubility of quartz is 6-10 ppm, cristobalite 20-30 ppm, and amorphous silica 60-130 ppm (Williams *et al*, 1985). The solubility of all forms of silica increases dramatically with temperature with thermodynamic calculations indicating values of 700 and 1800 ppm for amorphous silica at 100°C and 350°C respectively. In short, the solubility of glass shards within pore fluids of pyroclastic flows will be 2 orders of magnitude greater than concentrations accompanying pressure solution and formation of quartz overgrowths in "normal" sandstones. In addition, the rate of these

dissolution reactions will be exponentially increased at the higher temperatures according to the Arrhenius equation.

In the Garth and Racks Tuff in the Capel Curig anticline, the siliceous nodules common within the middle to upper portions of the non-reworked ash flow tuffs have been attributed to local upwards ascent of water vapour trapped beneath the ash flow tuff (Francis & Howells, 1973). As seawater only contains 2-14 ppm SiO_2 (see Williams & Crerar, 1985), the most logical source for silica contained within the quartz nodules is within the ash flow itself. As there are few pumice fragments within the Garth or Racks tuff from which to exsolve magmatic volatiles, it is suggested that the silica was derived from solution of glass shards in lower portions of the ignimbrite. As the ash flow tuff cooled from its base to seawater temperature, a water/steam reaction front would migrate upwards into the flow. Most silica was probably dissolved at temperatures less than 100°C when aqueous H_4SiO_4 species could form. Silica that was liberated would be transported upwards to the top of the flow by convecting water vapour cells to form the nodules rather than reprecipitated locally as in "normal" diagenesis. Thisprocess would represent the "additional chemical or physical agent" suspected by Howells & Leveridge (1980) to explain why deformation of glass shards "takes place at a lower temperature" than the 550°C threshold (Fisher & Schmincke, 1984, p. 213) that governs subaerial welding. If this model is correct, and significant changes in the morphology of glass shards occurred by temperature-enhanced pressure solution, rather than plastic deformation, than it is dubious whether one should even refer to these diagenetic vitroclastic textures in subaqueous deposits as welding (*sensu* Smith, 1960; Fisher & Schmincke, 1984). As noted many years ago by Pettijohn *et al* (1973, p. 44), there is probably only a "semantic difference between consideration of a welded tuff as a volcanic rock or as a sediment in which the conditions for **cementation** are especially

related to temperature. By analogy with processes responsible for oil maturation, one can envisage a time-temperature window for deformation of glass shards in hot subaqueous pyroclastic flows, with the relative importance of plastic to diagenetic deformation increasing at higher temperature and in anhydrous conditions. The survival of brachiopods in such a fluid regime is not a problem as carbonate solubility decreases with increasing temperature (ibid, p.422); in fact, the local presence of these carbonates and related increase in pH acts as a catalyst for the silica diagenetic reactions (Williams & Crerar, 1984).

7.3.3.2 IMPLICATIONS ON WELDING FROM NORTH WALES

All ash flow tuffs in North Wales were disrupted upon entry into the sea, and only retained enough heat to weld for short distances (less than 5 km) past the shoreline. The only evidence for the development of welding textures at water depths greater than flow thickness is within the Capel pods which are now interpreted as small remnants of a much larger pyroclastic flow. In this light, it is necessary to ask whether any ash flow tuffs can weld subaqueously. A preliminary review of literature (Fig. 7.82), assuming that the welding is not diagenetic, metamorphic, or transported as a plug into marine environments, suggests that development of welding is restricted to either extremely shallow water (less than flow thickness, *eg.* North Wales; Devonian, Germany, Scherp & Grabert, 1981) from flows entering the sea or to deep water (greater than 500 m, *eg.* Ordovician, South Wales, Kokelaar *et al*, 1985) where hydrostatic pressures confines submarine eruption columns and the increased solubility of water in glass decreases its viscosity (Sparks *et al*, 1980). Welded tuffs would be most unlikely at intermediate water depths (*ca.* 50-500 m). Here, eruption columns would mix considerably with water resulting in cool volcaniclastic debris flows or high density turbidity currents (*cf.* Fiske & Matsuda, 1964); with respect to ash flows entering the sea from land, water will continually be ingested into the flow along its turbulent base and front;

deflections from topographic barriers would result in catastrophic incorporation of significant amounts of additional water through flow transformations (Fisher, 1984). Any welding surviving these processes will be patchy and confined to 'hot spots' which ingested less water (*eg.* Capel pods; Palaeogene, Phillipines; Fernandez, 1969). More extensive welding at intermediate depths would only occur in special cases (*eg.* Yr Arddu tuffs, Howells *et al*, 1987) where the tuff is laterally restricted and the eruption column essentially falls back onto the vent. It is dubious, however, whether these laterally-confined deposits should be called ash <u>flow</u> tuffs. The higher proportion of subaqueous welded tuffs in the Archean is attributed to its hotter and structurally weaker (*ie.* low fountaining eruption columns) crust in comparison with the Phanerozoic.

7.4 CONTROLS ON VOLCANIC ERUPTIONS

Primary pyroclastic rocks in North Wales were largely derived from explosive fragmentation of rhyolite magmas. Major and trace elements (Fig. 7.84) indicate a high-K subalkaline chemical affinity (Leat *et al*, 1986; Fig. 7.83), with most samples containing concentrations of SiO_2 greater than 73%, TiO_2 concentrations less than 0.35% and Zr ranging from 150-350 ppm. Within the Capel Curig Volcanic Fm. (CCVF), the concentrations of Zr, Nb, Y and TiO_2 increase with each successive eruption episode (*ie.* member); a similar pattern has been observed within the Lower Rhyolitic Tuff Fm (LRTF). It would be tempting to regard the chemical trend within the entire CCVF to compositional zonation produced within a chemically zoned magma chamber. As noted by Leat *et al* (1986), however, the enrichment of Zr and Y in the <u>later</u> eruption products is opposite to trends observed in all other chemically zoned ash flow sheets (Hildreth, 1981); he resolved this conundrum by suggesting that after each eruption phase of the Snowdon volcanic centre, magma remaining in the high-level magma chambers evolved towards more differentiated compositions. The genesis of the CCVF is more complex. Any

models need to consider: 1) the substantial time periods occurring between eruption of the various volcanic members (1, 2 & 3/4), 2) that there were several volcanic vents, and 3) that the volcanic centre progressively stepped to the SW throughout the formation. Trace element data (Fig. 7.84) suggest that two broad groups of magma can be recognised within the CCVF, corresponding to the 1st/2nd and 3rd/4th volcanic members. These are represented on the plot of Nb vs Zr by two straight-line log-log trends at Nb/Zr ratios of 0.097 and 0.058, which might correspond to the Ralyeigh fractional crystallisation law (Ball & Merriman, 1987). The two groups are thought to reflect the independent evolution of two separate high-level magma chambers (Fig. 7.85). If subsidence patterns following eruptions can be related to magma migration as suggested (Chapt. 6.3), then this would imply connection of magma as a sill at mid-crustal levels.

Setting aside these intriguing petrogenetic and magma-plumbing problems, the differences in chemical composition within the CCVF and between the various formations, with the total variation in absolute SiO_2 being less than 2%, is insignificant when compared with broad range of eruptive styles and result pyroclastic rocks. The Garth and Racks Tuff, for example, each resulted from a single volcanic event producing vitric-rich laterally extensive ash flow sheets. The eruption and emplacement of the Pitts Head Tuff was similar with differences being a more prolonged eruption (?an open fountaining vent), concurrent volcanotectonic collapse at source, and pumice as a common product. All these volcanic units contrast with the 4th Mbr. of CCVF and the LRTF which indicate a series of small closely spaced eruptions producing fine- to very coarse-grained volcanic ash with varying crystal/lithic components.

In absence of compositional variation, the contrasting nature of the explosive pyroclastic eruptions will be governed by: (1) viscosity: depending on volatile content, temperature, and amount of bubbles and

crystals, (2) type and amount of magmatic volatiles, and (3) rate of ascent and depth of vesiculation of magma. It would be naive to expect the type of pyroclastics produced to be related to only one of these interdependent variables. With a single eruption cycle, the amount of magmatic volatiles was proabably most important. The changing eruptive behaviour, for example, within the 3rd-4th volcanic member of CCVF succession or bipartite "tuffite" units can be attributed to a stratified magma chamber with explosive eruption of the volatile-rich cupola followed by less explosive ejection of some of the remaining volatile-depleted magma containing phenocrysts. The coeval eruption of these two components in the 3rd and 4th members might be explained by the control of topography on the level at which volcanic conduits access the magma chamber (Fig. 7.85). In larger volcanic eruptions, however, the crystal-rich component was never erupted bu emplaced as shallow-level intrusive to extrusive domes (*cf.* Campbell *et al*, 1987); neglecting size and cooling constraints, the position to which these ascend will depend on their rheology as determined by the competing influences of crystal content, volatile concentration and amount of vesiculated gas.

When one attempts to understand the differing styles exhibited by separate volcanic centres through basin evolution, however, and takes into account (1) the magnitude (volume) of the volcanic event, (2) the size and variability of the individual eruptive phases, (3) the eruption duration and recurrence interval, and (4) extent of associated volcanotectonic movements, one gets the distinct impression that the strength of the continental crust is also involved. The arguments presented in the previous chapter can be extended as follows. With eruption of each member of the CCVF, the volcanotectonic movements progressively fractured the upper crust; the effective reduction in lithostatic load and confining pressure limited the amount of exsolved volatiles which could be contained at the top of the magma chamber. Eruptions became more frequent and less explosive. The Pitts Head Tuff was

erupted from a "virginal" terrain which had suffered few previous volcanic eruptions; the high confining pressures of the lithostatic load enabled development of a thick accumulation of vesiculated magma at the top of the magma chamber which was eventually erupted in a single outburst. The eruption of the Snowdon Volcanic Group represents the last expression of Ordovician volcanicity in North Wales. As shown by the Lower Rhyolitic Tuff Fm., or Crafnant Volcanic Fm., the continental infrastructure had suffered so many "ups and downs" that a series of small eruptions was inevitable. The leaky crust also allowed basaltic magma, which had had always been an underlying control of basin evolution (see below), to finally leave its signature in the stratigraphic column with the climatic eruptions Bedded Pyroclastic Formation.

As an epilogue to this last analytical chapter, it is pertinent to ask what triggered the large rhyolitic volcanic eruptions? Some light is cast on a viable mechanism by: 1) the occurrence of sandstone rich unusually rich in basaltic detritus, and derived from the west, beneath the PHT on Moel Hebog (Chapt. 4.3.5.3) and 2) the extrusion of basalts along the Glyder Fawr fracture immediately prior to eruption of the LRTF from nearby vents along LLanberis pass (Chapt. 7.2.7, Fig. 5.33). It is suggested that rhyolitic eruptions were triggered according to the model of Sparks *et al* (1977) by superheating and rapid convection induced by mixing with small amounts of basalt which was introduced into magma chambers during episodes of crustal extension.

Chapter 8

SEDIMENTATION, TECTONICS & VOLCANICITY: ORDER IN CHAOS

" The Lord may be subtle, but malicious He is not "

Albert Einstein

8.1: PLATE TECTONIC SETTING: A BACK-ARC TO THE TORNQUIST.

Before synthesizing the interplay between sedimentation, tectonics, and volcanicity into a unified theory, it is necessary to comment on the plate tectonic setting of North Wales and the Welsh basin during the Ordovician. It has previously been assigned to a back-arc basin (Dewey, 1969; Mitchell & Reading, 1970; Phillips *et al*, 1976; Kokelaar, 1984) related to subduction of the Iapetus lithosphere. Although an origin as an ensialic marginal basin appears certain, the relationship to the Iapetus ocean is questionable. If one examines the tectonic fabric which governed the formation, distribution, and preservation of the volcanic and sedimentary rocks one begins to wonder if the axis of the basin in North Wales was actually aligned NW-SE (Fig. 8.1 a,b). The dominant structural weaknesses which controlled basin evolution were the Conwy, Llanberis Pass and to a lesser extent Cwm Pennant fault systems. Although all fault orientations controlled the emplacement of late-stage volcanic plutons (see Campbell *et al*, 1987), the large rhyolitic eruptions such as the Garth and Racks tuff, Pitts Head Tuff and Lower Rhyolitic Tuff were all sourced by volcanic vents and often preceded by basaltic eruptions from the NE-SW lineaments. The eastern margins of the Pitts Head Tuff volcanotectonic trough at Llwyd Mawr, and the asymmetric (?half-graben) Lower Rhyolitic tuff "caldera" were controlled by the same fractures. With regards to sedimentation, coarse-grained alluvial fans which filled the two main depocentres about Beddgelert and Tryfan reflected uplift on these NW-SE master faults. Braidplains beneath Capel Curig Volcanic Fm. were structurally bound by the Conwy and Llanberis Pass fault systems. The latter fault was later used to

accommodate increased amounts of structural shortening to NE of Snowdon Massif during Caledonide tectonic movements, and can be recognised today by a distinctive aeromagnetic signature (?Tertiary dykes) within the Irish sea (Fig. 8.2). The more diffuse NE-SW lineaments are often structurally paired (*eg.* Menai Straights system, Nant Peris/Idwal-Hebog faults, Nantmor-Cowlyd/Yr Arddu faults) and probably originated as strike-slip faults during latest Precambrian accretion of the various suspect terrains which eventually became Avalonia. Their effect on the Welsh structural fabric was increased during latest Silurian to Devonian closure of the Iapetus (Woodcock, 1984. by minor strike-slip movements. However, their only influence in the Ordovician Period, however, was to act as sources for minor volcanic eruptions, to occasionally feed miniscule alluvial fans, or more importantly to passively allow differential subsidence and/or uplift of various crustal segments along the axis the basin, as dictated by movement of magma at depth.

Kokelaar (1988, in press) has demonstrated how the NNW-SSE trending swarm of sheet-like intrusions at Rhobell Fawr (Kokelaar, 1979), which include a 1 km wide zone of 100% dykes, signifies substantial degrees of EAST-WEST extension. An analogous but smaller-scale set of NW-SE trending dykes in North Wales occurs associated with pillow basalts at Glyder Fawr (Chapt. 7.2.7). Kokelaar (1988) invokes a rather tenuous correlation of the parallel Trawfynydd-Rhobell Fawr and Conwy fractures using a previously undiscovered array of minor faults to explain relationships at Rhobell Fawr by N-S sinistral transtension associated with the currently fashionable models for closure of the Iapetus using strike-slip tectonics (*eg.* Soper & Hutton, 1984). The results of this study, however, suggest that it is simpler to abandon the existing polarity for the formation of the Welsh basin and accept the ENE-WSW extension at its face value. As a natural extension of the ideas formulated in this thesis, Orton & McKerrow (in prep)

have shown through an integration of available sedimentological, palaeontological, volcanic and tectonic data from all parts of the Avalonian microcontinent, that the Welsh marginal basin was in fact probably related to SW subduction of the Tornquist under the Midland platform (Fig. 8.3). A corollory of this model is that their was NO subduction of the Iapetus under the southern British Isles. Later reactivation of this Tornquistian fabric as a splaying network of shear structures controlled offset patterns developed during propagation of the North Sea Rift and resulted in an en echelon series of rhomboidal pull-apart basin (Cartwright, 1987, 1988).

8.2: VOLCANIC CONTROL ON SEDIMENTATION

Most research on volcaniclastic sedimentation has been from convergent margin volcanic arcs (*eg.* Mt. St. Helens), areas of high relief and common andesite volcanism (Mathisen & Vondra, 1983; Fritz & Harrison, 1985; Smith, 1986). In these terrains, non-marine sedimentation proceeds directly in unison with volcanism as a series of relatively short-lived pulses of aggradation (20-30 years) separated by longer periods (80-125 years) of comparatively minor volcanicity and little resultant sedimentation (Vessell & Davies, 1981). The development of large-scale coarsening-upward sequences is usually thought to reflect increasing **constructional topography** (*ie.* growth of stratovolcano) in the volcanic arc (Fritz & Harrison, 1985; Smith,1987), rather than the amount of differential **tectonic uplift** between fault-bound highlands and depositional basins (Miall, 1981; Nilsen, 1984). The wedge-shaped aprons produced by volcanism-induced aggradation will taper away from the volcanic axis and can attain thicknesses of up to several thousand metres. Proximal to distal facies changes reflect distance from the magmatic arc with deposition of lavas and primary pyroclastics in the volcanic core, debris flows (lahars) and hyperconcentrated flood flows in medial areas, and braided or meandering stream facies distally (Vessell & Davies, 1981).

Good modern examples of convergent margin volcanic arcs are provided by Tenerife in the Canary Islands and Santorini, Greece. Primary volcanics, both pyroclastics and lavas, dominate depositional successions on both islands (Fig. 7.65, Fig. 8.4-8.8). The contrasting amounts of intercalated epiclastic component between and even on (*eg*. Tenerife) these volcanic islands can be related to climate with a larger proportion of volcanic ash reworked and redeposited by "everyday" alluvial processes in more humid terrains. With development of radial drainage patterns of limited catchment area around volcanic cones, comparatively high volcanic relief, and frequent reorganisation of depositional systems by volcanic events, debris flows and hyperconcentrated flood flows will be dominant mechanisms responsible for redistribution of volcaniclastics. Caldera fill-successions (Fig. 8.9), surprisingly, are one the best areas for preservation of epiclastic transport processes.

In the United Kingdom, Rhobell Fawr in Wales (Fig. 8.10) or more spectacularly the Lake District in Cumbria typify convergent margin volcanic arcs, both in the frequency of volcanic eruptions, and type of derivative epiclastic component (Fig. 8.11-8.12). The contrasting depositional style in Snowdonia cannot be related to either climate or distance from source. Both regions experienced a humid, temperate to subtropical climate; in both cases, epiclastic successions preserved are located within a few kilometres of probable volcanic vents. One must therefore look to variables such as magma chemistry, eruption size and frequency, grain size of pyroclasts produced and relative importance of intrabasinal volcanic and tectonic relief. In North Wales, depositional facies suggest gentle constructional topography and little intrabasinal volcanic relief. Individual pyroclastic eruptions, usually rhyolitic were energetic with long (100- 500,000 years) recurrence intervals. Large volcanic events, in producing welded sheets of ash flow tuff, obliterated pre-existing source terrains; any loose volcanic

ash resulting from smaller pyroclastic eruptions was usually contained within localised volcanic cones and had little effect on the surrounding sedimentary budget. Depositional facies of both eruption styles suggest little intrabasinal volcanic relief. As a result, pyroclastic units are laterally extensive and can be correlated geochemically to provide a useful stratigraphic framework for examination of the intercalated epiclastic successions.

Although over 1000 metres of such detritus are preserved from a wide variety of environments, few episodes of sedimentation directly resulted from penecontemporaneous volcanic activity. The middle portion of the Lower Moel Ddu. member (Chapt. 4.3.6.2) represents the only subaerial example, with a thick succession of fine-grained sheetfloods recording rhyolitic volcanism and coarse-grained hetrolithic sandstones deposited in braided channels or incision of meandering gullies reflected volcanically "quiet" periods (compare with Walton, 1979, 1986). Subaqueous examples are more abundant and include several of the sandstone lobes within the Upper Cwm Eigiau Fm.; the apparent dominance of volcanically-related sedimentation in this stratigraphic unit, however, is misleading in that the thick successions of unstudied cleaved mudstones were probably eroded by epiclastic processes. To generalise, most volcaniclastic sedimentation represents reworking of previously deposited pyroclastics by organised depositional systems during the long periods of time between volcanic events. Sandstone petrography indicates derivation of most detritus from uplifted fault blocks of volcaniclastic sediment, rhyolitic plutons, or granite rather than penecontemporaneous pyroclastics. The dominant control of volcanism on sedimentation was in causing numerous tectonic changes within the basin, with the most significant changes in basin palaeogeography related to silicic eruptions. Magmatically driven tectonic changes will of course be most feasible and most visible in extensional basins (see

citations in Fig. 6.1). In compressive situations, fault blocks will be structurally coupled making relative movements more difficult; in addition, if uplift does occur it would only make existing stratovolcanoes "a bit bigger" and not result in any profound changes in basin configuration and palaeogeography.

What is even more remarkable, is that when the volcanological, sedimentary, tectonic and even palaeontological data are fully intergrated, an uncanny harmony energes from the chaos that one would intuitively suspect for a volcanically influenced region. This can be seen at a complete hierarchy of scales. At the most detailed level of observation, for example, the amount of basin subsidence and development of either subaerial or subaqueous (lacustrine) environments on the Moel Ddu alluvial plain and magnitude of discharge of fine-grained volcanic ash were both determined by the rate of magma supply. At a moderate scale, the progradation of deltaic lobes was either enhanced through increased availability of sediment (*eg.* Capel Curig Mbr.) or more commonly completely interupted (*eg.* Gwern Gof, Moel Hebog) by tectonic events associated with volcanicity. Finally, at a regional level of examination, tectonic changes resulting from the emplacement, migration and vesiculation of magma bodies parental to the large volcanic eruptions determined both basin configuration and its direction of infill; the influx of oxygenated marine water into the basin and reduced sediment supply during subsidence following these large eruptions, when combined with added nutrients and a warm substrate, provided conditions ideal for the proliferation of shelly fauna; sedimentary succession directly overlying the Garth Tuff (*eg.* Capel Curig anticline) or Pitts Head Tuff (*eg.* Pont-y-Gromlech) contain a more substantial proportion of brachiopods than found elsewhere in strata of comparable depths. A progressive fracturing of the continental crust, resultant reduction in its strength, and tranformation with time in the magnitude and style of large

volcanic eruptions was a natural result of these large, basin-wide, volcanotectonic events; the inability to maintain large volatile-rich magma bodies at upper crustal levels aided the gradual subsidence of the basin beneath sea level during the Longvillian eustatic transgression despite continued volcanic activity.

Although the evidence in favour of such magmatically-driven tectonic movements in continental crust is limited at present the profound implications for the style of basin subsidence, nature and distribution of volcaniclastic sedimentation, geothermal gradients, and patterns of hydrothermal circulation during early phases of continental rifting warrant further investigation of the problem. Most continental rifts (*eg.* Baikal, East Africa, North Sea, Red Sea, Rhine graben, Rio grande) do show evidence for some degree of accompanying volcanism. Along the Brazilian continental margin Early Cretaceous igneous activity was associated with rifting of Gondwana (Ojeda, 1982) with extrusion of basalt/andesite into the developing half-grabens; likewise, Albian-Cenomanian pyroclastic activity in paralic to shallow marine and fluviatile sequences characterizes the earliest rift sedimentation within the Benue Trough in west Africa (Petters, 1978). As shown by Cohen *et al* (1986), a spectrum of rift models can be postulated based on the degree of volcanism relative to rift subsidence. In North Wales, the volcaniclastic sedimentation illustrates the strong control of this volcanism on subsidence at even a local scale.

Recent research on continental rifts and development of sedimentary basins has concentrated on large scale geodynamic aspects with much attention focussed on mathematical modelling involving either homogenous stretching of the whole lithosphere (McKenzie, 1978) or displacement on lithosphere scale shear zones (Wernicke, 1985; Coward, 1986). These simplistic models, however, have had limited success in accounting for observed patterns of subsidence and many authors have found it necessary to

invoke mechanism that in effect introduce more heat into the system such as dyke intrusion and two layer stretching (*eg.* Royden & Keen, 1980; Sclater *et al*, 1980). In a thought-provoking review of the fracture mechanism of magma transport from the mantle to the surface Shaw (1980) has noted that: (1) if pore-fluid pressure approximates gravitational load, magma injection by extensional fracture can occur at any depth in the mantle, (2) fluid pressure from melting reactions, which produces a 10% increase in volume for most rocks, acts in the asthenosphere in the same fashion as hydrological formation pressure which governs development of gravity sliding, decollement and growth faulting in near-surface sedimentary successions and (3) through a process of interactive feed-back relations the distribution of magma regulates plate-tectonic motions over the earth. The existence of this mobile fluid phase, not only explains the varying subsidence patterns of continental rifts, but solves many "space" problems arising from models which invoke development of brittle failure (*eg.* Wernicke, 1985) and pull-apart morphologies (Cartwright, 1988) in the lower crust associated with major lineaments which transect the entire lithosphere.

In North Wales, if one accounts for the influence of crustal strength on eruption frequency, then a surprising quasi-deterministic temporal order can be established for the evolution of the basin; the basal portion of the Llewelyn Volcanic Group, Capel Curig Volcanic Formation, Pitts Head Tuff Fm., and Snowdon Volcanic Group are all of comparable volumes and separated by comparable periods (*ca.* one million years) as inferred (Fig. 6.3) from thickness of epiclastic successions. These rhythmic outbursts of volcanicity must reflect periodicity in the rate of magma emplacement into the upper continental crust. This begs the question whether magma emplacement results from, modifies, or even determines the tectonic stress regime with the continental crust. It is recognised that the major volcanic eruptions in North Wales (*eg.* PHT, LRTF) were triggered by episodes of

crustal extension (Chapt. 7.4). But what causes continental plates to move and rifts to form? It is suggested here, that migration and forceable injection of magmatic fluids into fractures is a cause rather than a result of the propagation of continental rifts; this is supported by a recent analysis of plate tectonic processes in the Long Valley area, California, where geophysicists (Julian & Sipkin, 1987) have related earthquake mechanisms to tensile failure from magma intrusion rather than shear faulting.

8.3 SO WHAT?

> "A wise writer will feel that the ends of a study and composition are best answered by announcing undiscovered regions of thought and so communicating through hope, new activity to the torpid spirit"
>
> *Emerson, 1836, Nature.*

Returning to a less philosophical mode of thought, it is pertinent to ask what we have learned, possible fruitful area of future research, and why study volcaniclastics. If the author could leave the reader with one theme, it would be that in a volcanically-influenced setting one cannot isolate the volcanism, sedimentation, palaeontology, or tectonics as separate areas of study. Full comprehension of the dynamics of the depositional basin can only be achieved by a complete integration of all these aspects. It is hoped that this "case study" of volcaniclastic sedimentation in an extensional marginal basin has illustrated the necessity of such an approach.

It would be ludicrous to presume that we now understand volcaniclastic sedimentation in continental rifts, or even the sedimentology of North Wales. The results presented have merely scratched the surface of a 'classic' area of British geology, and interpretations presented are limited by the rather meagre data base from which they were derived. Many of the outstanding problems regarding the origin and development of rifts, the volcanic influence on sedimentation, and even erection of facies models

applicable to non-volcanic settings can be addressed through further study of Ordovician successions in Wales. A representative selection of useful topics is tabulated in Figure 8.14 in hope of spurring future research into a facet of British geology still neglected by natives.

Upon completion of this study, the almost complete absence of sedimentologists from the annals of Snowdonian geological research is even more of a mystery. The pages of photographs illustrate that sedimentary structures abound in greater proportion than found within more mundane Upper Palaeozoic or Mesozoic sequences. The area has been continually visited by field groups since the 19th century; satisfactory geological maps have existed for substantial proportions of the region for 30 years. It would seem that the only possible reason is a phobia of "volcanoes". In an analysis of controls on deltaic sedimentation Orton (1988; enclosure 4) down-played the volcanic component to avoid scaring sedimentologists away. The cynic might argue that the study of volcaniclastics has almost no relevance to the bulk of sedimentology and problems experienced in "normal" sedimentary basins. It is hoped that upon reading this thesis, the reader will agree that the converse is true. The chronostratigraphic markers that exist throughout volcanically-influenced settings provide a unique framework in which to examine all processes of physical sedimentology and the formation of sedimentary basins in general. The continual variability in grain size within volcaniclastic successions gives an unparalleled handle on the control of this variable on both the facies and depositional environments that will develop, as illustrated by nearshore sequences. Whereas in "normal" successions the slower rates of sedimentation lead to frequent punctuation of the succession, in volcaniclastic terrains one has a comparatively complete depositional record, with fewer hiatuses.

REFERENCES

AIGNER, T. (1985). Storm depositional systems: dynamic stratigraphy in modern and ancient shallow marine sequences. Springer Verlag, Berlin.

ALAM, M.M., CROOK, K.A.W. & TAYLOR, G. (1985). Fluvial herringbone cross stratification in a modern tributary mouth bar, Coonamble, New South Wales, Australia. Sed., 32 235-244.

ALLEN, J.R.L. (1971) Mixing at turbidity current heads and its geological implications. J. Sediment. Petrol. , 41, 97-113.

ALLEN, J.R.L. (1983). Studies in fluviatile sedimentation: bars, bar complexes and sandstone sheets (low sinuosity braided streams) in the Brownstones (L. Devonian), Welsh Borders. Sed. Geol. 33 237-293.

ALLEN, J.R.L. (1984) Sedimentary structures. Their character and physical basis. 2nd edition., Elsevier, Amsterdam.

ALLEN, P.A. (1981). Some guideline for reconstructing ancient sea conditions from wave ripple marks. Marine Geology, 43, 59-67.

ALLEN, P.A. (1985). Hummocky cross-stratification is not produced purely under progressive gravity waves. Nature, 313, p.562-564.

ANDRE, L., HETROGEN, J., & DEUTSCH, S. (1986) Ordovician-Silurian magmatic provinces in Belgium and the Caledonian orogeny in middle Europe. Geology, 14, 879-882.

BACHMAN, G.O. & MEHNERT, H.H. (1978). New K.Ar dates and late Pliocene to Holocene geomorphic history of the central Rio Grande region, New Mexico. Bull. Geol. Soc. Am. 89, 283-292.

BAKER, V.R. (1977) Stream channel responses to floods , with examples from central Texas. Bull. Geol. Soc. Am., 88, 1057-1071.

BALL, T.K. & MERRIMAN, R.J. (1987) The Petrology and Geochemistry of the Ordovician Llewelyn Volcanic Group, Snowdonia, North Wales. British Geological Survey, Mineralogy and Petrology Report 87/21

BANCROFT, C. (1933) Correlation tables of the Stages Costonian to Onnian in England and Wales. Blakeney, Glos., p.1-4.

BANCROFT. (1945). The brachiopod zonal indices of the stages Costonian to Onnian in Britain. J.Paleont., 19, 181-252.

BARKER, D.S. (1983) Igneous rocks. Prentice-Hall, New Jersey, 417pp.

BARRELL, J. (1912) Criteria for recognition of ancient delta deposits. Bull. Geol. Soc. Am., 23, 377-446.

BASSETT, D.A. (1963) The Welsh Palaeozoic Geosyncline: a review of recent work on stratigraphy and sedimentation. In: "The British Caledonides" (ed. by M.R.J. Johnson and F.H. Stewart), pp. 35-69.

BASSETT, M.G. (1984) Lower Palaeozoic Wales - a review of studies in the past 25 years. Proc. Geol. Assoc., 95, 291-311.

BEAVON, R.V. (1963). The succession and structure east of the Glastyn

river, north Wales. J.Geol.Soc. London, 119, 479-512.

BEAVON, R.V. (1974) Discussion of Ordovician transgressive ash flow tuffs of Snowdonia, North Wales. J. Geol. Soc. London, 130, 479-480.

BEAVON, R.V. (1980). A resurgent cauldron in the early Palaezoic of Wales, U.K. Journal of Volcanology and Geothermal Research 7, p. 157-74.

BEAVON, R.V., FITCH, F.J. & RAST, N. (1961) Nomenclature and diagnostic characters of ignimbrites with reference to Snowdonia. Liverpool and Manchester Geol. J., 2, 600-611.

BELT, E.S. & BUSSIERES, L. (1981) Middle Ordovician submarine fans and associated facies, northeast of Quebec City. Can. J. Earth Sci., , 18, 981-994.

BENTON, M.J. & GRAY, D.I. (1981). Lower Silurian distal shelf storm-induced turbidites in the Welsh Borders: sediments, tool marks and trace fossils. J. geol. Soc. London, 138, 675-94.

BERGH, S.G. & TORSKE, T. (1986) The Proterozoic Skoadduvarri sandstone formation, Alta, Northern Norway: A tectonic fan delta complex. Sed. Geol., 47, 1-25.

BERGMAN, K.M. & WALKER, R.G. (1987). The importance of sea level fluctuations in the formation of linear conglomerate bodies; carrot Creek Mbr of Cardium Fm, Cretaceous western Interior seaway, Alberta, Canada. J. Sediment. Petrol., 57, 651-665.

BERRY, W.B. & WILDE, P. (1978). Progressive ventilation of the oceans an explanation for the distribution of the lower Paleozoic block shales. Am. J. Sci. 278, 257-275.

BEVERAGE, J.P. & CULBERTSON, J.K. (1964) Hyperconcentrations of suspended sediment. J. Hydraulics Division. Proc. Am. Soc. of Civil Engineers, 90 , HV6, 117-128.

BEVINS, R.E. (1982). Ordovician igneous activity in southwest Dyfed. In: Bassett,M.G. (ed.) Geological Excursion in Dyfed, South-West Wales, 65-80. National Museum of Wales, Cardiff.

BEVINS, R.E. & ROWBOTHAM, G.(1983) Low grade metamorphism within the Welsh sector of the paratectonic Caledonides. Geol. J., 18, 141-167.

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., 95, 337-347.

BHATTACHARYYA, D.P. & LORENZ, J.C. (1983). Different depositional settings of the Nubian lithofacies in Libya and southern Egypt. In: IAS6, p.435-448.

BJORLYKKE, K. (1971). Petrology of Ordovician sediments from Wales. Norsk geol. Tidsskr. 51, p.123-139.

BJORLYKKE, K. (1974). Geochemical and mineralogical influence of Ordovician island arcs on epicontinental clastic sedimentation. A study of lower palaezoic sedimentation in the Oslo region, Norway. Sedimentology v21

p.251-272.

BLATT, H. (1985) Provenance studies and mudrocks. J. Sediment. Petrol., 55, 69-76.

BLOOS, J. (1982) Shell beds in the Lower Lias of South Germany - facies and origin. In: Cyclic and event stratification (ed. by G. Einsele and A Seilacher), Springer-Verlag, Berlin, p. 223-239.

BOOTHROYD, J.C. (1972) Coarse grained sedimentation on a braided outwash fan, northeastern Gulf of Alaska: Technical Report No. 6-CRD, Coastal Research Division, Univ. of South Carolina, Columbia, S.C. 127pp.

BOOTHROYD, J.C. & ASHLEY. G.M.(1975) Processes, bar morphology and sedimentary structures on braided outwash fans, NE Gulf of Alaska. In: Glaciofluvial and glaciolacustrine sedimentation (Ed. by A.V Jopling and B.C. McDonald). Spec. Pub. Soc. Econ. Paleont. Min., Tulsa, 23, 193-222.

BOOTHROYD, J.C. & NUMMEDAL, D. (1978) Proglacial braided outwash: a model for humid alluvial fan deposits. In: Fluvial sedimentology (Ed.. by A.D. Miall). Can. Soc. Petrol. Geol. Mem. 5, 641-668.

BOSWELL, P.G.H. (1960) The term greywacke. J. Sediment. Petrol., 30, 154-157.

BOSWELL, P.G.H. (1961) The case against a Lower Palaeozoic geosyncline in Wales. Lpool and Manchr. geol. J., 2, 612-624.

BOURGEOIS, J. (1980) A transgressive shelf sequence exhibiting hummocky cross stratification: The Cape Sebastian sandstone (Upper Cretaceous), SW Oregon. J. Sediment. Petrol., 50, 681-702.

BOURGEOIS, J. & LEITHOLD, E.L. (1984). Wave reworked conglomerates-depositional process and criteria for recognition. In: Sedimentology of gravels and conglomerates. (ed. by Emlyn. H. Koster & R.J. Steel). Can. Soc. Petrol. Geol memoir 10 p.331-334.

BRAITHEWAITE, R.L. (1974) The geology and origin of the Rosebury Ore Deposits, Tasmania. Econ. Geol., 69, 1086-1101.

BRANNEY, M.J. (1986) Discussion: Isolated pods of subaqueous welded ash flow tuff: a distal facies of the Capel Curig Volcanic Formation (Ordovician), North Wales. Geol. Mag., 123, 589-590.

BRENCHLEY, P.J. (1964) Ordovician ignimbrites in the Berwyn Hills, North Wales. Geol. J., 4, 43-54.

BRENCHLEY, P.J. (1969) Origin of matrix in Ordovician greywackes, Berwyn Hills, North Wales. J. Sediment. Petrol. , 39, 1297-1301.

BRENCHLEY, P.J. (1969) The relation between Caradocian Volcanicity and sedimentation in North Wales. In: "The Precambrian and Lower Paleozoic Rocks of Wales" (ed. by A. Wood), Univ. of Wales, pp. 181-202.

BRENCHLEY, P.J. (1978) The Caradoc rocks of the north and west Berwyn Hills, north Wales. Geol. J., 13, 137-164.

BRENCHLEY, P.J. (1985) Storm influenced sandstone beds. Modern Geology, 9, 369-396.

BRENCHLEY, P.J. & NEWALL, G. (1980). A facies analysis of upper Ordovician regressive sequences in the Oslo region, Norway, a record of glacio euatatic changes. Paleogeogr. Paleoclimatol. Paleoecolog 31, 1-38.

BRENCHLEY, P.J. & NEWALL, G. (1982) Storm influenced inner shelf sand lobes in the Caradocian (Ordovician) of Shropshire, England. J. Sediment. Petrol., 52, 1257-1269.

BRENCHLEY, P.J. & PICKERILL, R.J. (1980) Shallow subtidal sediments of Soudleyan (Caradoc) age in Berwyn Hills, North Wales and their paleogeographic context, Proc. Geol. Assoc., 91, 177-194.

BRENCHLEY, P.J., NEWALL, G. & STANISTREET, I.G. (1979). A storm surge origin for sandstone beds in an epicontinental platform sequence, Ordovician, Norway. Sed.Geol. v22 p.185-217.

BRIDGES, P.H. (1976) Lower Silurian transgressive barrier island deposits, southwest Wales. Sedimentology, 23, 347-362.

BRITTER, R.E. & LINDEN, P.F. (1980). The motion of the front of a gravity current travelling down an incline. J. Fluid. Mech., 99, p.531-543.

CHAN, M.A. & DOTT, R.H., Jr., (1986). Depositional facies and progradational sequences in Eocene wave-dominated deltaic complexes, southwest Oregon: American Association of Petroleum Geologists, 70, p.415-429.

BRITTER, R.E. & SIMPSON, J.E. (1978). Experiments on the dynamics of gravity current head.

BROWN & FISHER. (1977). Seismic stratigraphic interpretation of depositional systems: examples from the Brazilian rift and pull apart basins. In: C.E. Payton (ed); Seismic Stratigraphy-Applications to hydrocarbon exploration, Am. Assoc. Petrol. Geol. Mem v26 p.213-248.

BUDDINGTON, A.F. & LINDSLEY, D.H. (1964). Iron titanium oxide minerals & synthetic equivelants. J. Sediment. Petrol. 5, 310-357.

BULL, W.R. (1962) Relation of alluvial fan size and slope to drainage basin size and lithology in western Fresno County, California. U.S. Geol. Survey Prof. Paper No. 450B.

BULL, W.R. (1972) Recognition of alluvial fan deposits in the stratigraphic record. In: Recognition of Ancient Sedimentary Environments (Ed. by J.K. Rigby and W.K. Hamblin). Spec. Pub. Soc. Econ. Paleont. Min., 16, 63-83.

BUSBY-SPERA, C.J. (1986) Depositional features of rhyolitic and andesitic volcaniclastic rocks of the Mineral King Submarine Caldera Complex, Sierra Nevada, California. J. Volcanol. Geotherm. Res., 27, 43-76.

CALLENDAR, D.L. & FOLK, R.L. (1958) Idiomorphic zircon: key to volcanism in the Lower Tertiary sands of central Texas. Am. J. Science, 256, 257-269.

CAMPBELL, C.V. (1966) Truncated wave ripple laminae. J. Sediment. Petrol., 36, 825-828.

CAMPBELL, C.V. (1976) Reservoir geometry of a fluvial sheet sandstone. Bull. Am. Assoc. Petrol. Geol., 70, 1009-1020.

CAMPBELL, S.D.G. (1983) The geology of an area between Bala and Betws-y-Coed, North Wales. Phd thesis (unpublished), Univ. of Cambridge.

CAMPBELL, S.D.G. (1984) Aspects of dynamic stratigraphy (Caradoc-Ashgill) in the northern part of the Welsh basin. Proc. Geologists Assoc., 95, 390-391.

CAMPBELL, S.D.G., REEDMAN, A.J., & HOWELLS, M.F. (1985) Regional variations in cleavage and fold development in North Wales. Geological Journal, 20, 43-52.

CAMPBELL, S.D.G., REEDMAN, A.J., HOWELLS, M.E. & MANN, A.C. (1987). The emplacement of geochemically distinct groups of rhyolites during the evolution of the lower rhyolitic tuff formation caldera (Ordovician), north Wales, U.K. Geol Mag., 124, 501-51.

CANT, D.J. & WALKER, R.G. (1976) Development of braided fluvial facies models for the Devonian Battery Point sandstone, Quebec. Can. J. Earth Sci., 13, 102-119.

CANT, D.J. & WALKER, R.G. (1978). Fluvial processes and facies sequences in the sandy south Saskatchewan river, Canada. Sedimentology, 25, 625-648.

CAREY, S.N. & SIGURDSSON, H. (1980) The Roseau Ash: Deep sea tephra deposits from a major eruption on Dominica, Leser Antilles Arc. J. volcanol. geotherm. Res., 7, 67-86.

CARSON, M.A. (1984). The meandering-braided threshold: a reappraisal. In: J. Hydrol, 73, p.315-334.

CARTWRIGHT, J.A (1988). A seismic interpretation of the Danish North Sea. Unpublished D. Phil thesis, Oxford Univerisity.

CARTWRIGHT, J.A. (1987). Transverse structural zones in continental rifts, an example from the Danish sector of the North sea. In: K.W. Glemmie & J. Brooks. (eds). 3rd conference of the Petroleum Geology of N.W. Europe.

CAS, R.A.F. & WRIGHT, J.V. (1987) Volcanic successions: modern and ancient: a geological approach to processes, products, and successions. Allen & Unwin, London, 528 pp.

CAVAZZA, W. (1986) Miocene sediment dispersal in the central Espanola basin, Rio Grande Rift, New Mexico, USA. Sedimentary Geology, 51, 119-135.

CAVE, R. (1965) The Nod Glas sediments of Caradoc age in North Wales. Geological Journal, 4, 279-298.

CHAN, M.A. (1985) Correlations of diagenesis with sedimentary facies in Eocene sandstones, Western Oregon. J. Sediment. Petrol., 55, 322-333.

CHAPIN, C.E. & LOWELL, G.R. (1979) Primary and secondary flow structures in ash-flow tuffs of the Gribbles Run palaeovalley, central Colorado. Geol. Soc. Amer., Sp. Paper 180, 137-154.

CHRISTIANSEN, R.L., LIPMAN, P.W., CARR, W.J., BYERS, F.M., ORKILD, P.P. & SARGENT, K.A. (1977). Timber Mt-Oasis Valley caldera complex of southern Nevada. Bull. Geol. Soc. Am., 88, 943-959.

CLIFTON, H.E. (1981) Progradational sequences in Miocene shoreline deposits, southeastern Caliente Range, California, J. Sediment. Petrol., 51, 165-184.

CLIFTON, H.E. & DINGLER, J.R. (1984). Wave formed structures and paleoenviromental reconstructions. Marine Geology., 60, 165-198.

CLIFTON, H.E., HUNTER, R.E. & PHILLIPS, R.L. (1971) Depositional structures and processes in the high energy nearshore. J. Sediment. Petrol., 41, 651-670.

CLIFTON,H.E. & HUNTER. R.E. (1986) Shelf and coastal facies response to eustatic sea level change in a rapidly subsiding Late Cenozoic Basin, San Francisco, California. In: Abstracts, 12th International Association of Sedimentologists Congress, 1986, Canberra, Australia. p. 62.

COCKS, L.R.M. & FORTEY, R.A. (1984) Faunal evidence for oceanic separations in the Palaeozoic of Britain. J. Geol. Soc. London, 139, 465-478.

COHEN, A.S., FERGUSSON, D.S., GRAM, P.M., HUBLER, S.I. & SIMS, K.W. (1986) The distribution of coarse grained sediments in modern Lake Turkana, Kenya: implications for clastic sedimentation models of rift lakes. In. Sedimentation in the African Rifts. (ed. by L.E. Frostick). Geological Society, Special Publication, 25, 127-139.

CORBETT, K.D.(1981) Stratigraphy and mineralization in the Mt. Read volcanics, western Tasmania. Economic Geology, 76, 209-230.

CORDELL, L., LONG, C.L. & JONES, D.W. (1985). Geophysical expression of the batholith beneath Questa caldera, New Mexico. Journal of Geophys. Res., 90, p.263-269.

COWARD, M.P. (1986) Heterogeneous stretching, simple shear, and basin development. Earth. Planet. Sci. Letters, 80, 325-336.

COWARD, M.P. & SIDDANS, W.B. (1979) The tectonic evolution of the Welsh Caledonides. In: "The Caledonides of the British Isles - reviewed.

COX, S.F. (1981) The stratigraphy and structural setting of the Mt.Lyell volcanic hosted sulfide deposits. Economic Geology, 76, 231-245.

CRIMES, T.P. (1970). A facies analysis of the cambrian of Wales. In: Paleogeogr. Paleoclimat. Paleocol. v7, p.113-170.

CUMMINS, W.A. (1957) The Denbigh grits: Wenlock greywackes in Wales. Geol. Mag., 94, 433-451.

CUMMINS, W.A. (1959). The Lower Ludlow Grits in Wales. Lpool & Manchr geol. J., 2, 159-179.

CUMMINS, W.A. (1962) The greywacke problem. Lpool and Manchr geol. J., 3, 51-72.

DABRIO, C.J. & POLO, M.D. (1988). Late Neogene fan deltas and associated

coral reefs in the Almanzora Basin, Almeria Province, south-eastern Spain. In: "Fan deltas- tectonic setting, sedimentology, and recognition" (ed. by. W. Nemec and R.J. Steel). Blackie, Glasgow. (in press)

DAILY, B., MOORE, P.S. & RUST, B.R. (1980). Terrestial marine transition in the Cambrian rocks of Kangaroo Island, south Australia. Sed. v27, p.379-399.

DAKYNS, J.R. & GREENLY, E. (1905) On the probable Pelean origin of the felsitic slates of Snowdon. Geological Magazine., 2, 541-549.

DAVIDSON-ARNOTT, R.G.D. & GREENWOOD, B. (1984). Bedforms and structures associated with bar topography in the shallow water wave enviroment, Kouchibouguac Bay, New Brunswick, Canada. J. Sediment. Petrol., 44, 698-704.

DAVIES, D.A.B. (1936) The Ordovician rocks of the Trefriw district (North Wales). Q. J. Geol. Soc. London, 92, 62-90.

DAVIES, D.K. (1979) Fluvial transport and downstream modification in an active volcanic region. IN: " Fluvial sedimentology" (Ed.. by A.D. Miall). Can. Soc. Petrol. Geol. Mem. 5, 61-84.

DAVIES, D.K., ALMON, W.R., BONIS, S.B. and HUNTER, B.E. (1978) Deposition and diagenesis of Tertiary-Holocene volcaniclastics, Guatemala. In: Aspects of diagenesis (ed. by P.A. Scholle and P.R. Schluger) pp. 281-306 SEPM sp. Pub. 26, 281-306.

DAVIS, R.A. (1986). Origin of barrier islands: some alternative ideas with examples from the Gulf coast of Florida, U.S.A. IN: Abstracts, 12th International Association of Sedimentologists Congress, 1986, Canberra, Australia. p.75.

DECKER, J. & HELMOLD, K.P. (1985) The effect of grain size on detrital modes: a test of the Gazzi-Dickinson point counting method. Discussion. J. Sediment. Petrol., 55, 618-620.

DEER, W.A., HOWIE, R.A. & ZUSSMAN, J. (1966) An introduction to the rock-forming minerals. Longman. 526 pp.

DENLINGER, R.P. & RILEY, F. (1984). Deformation of Long Valley caldera, Mono County, California from 1975-1982. J.G.R., 89, 8303-8314.

DENNY, C. (1967) Fans and pediments. Am. J. Science, 265, 81-105.

DEWEY, J. (1969) Evolution of the Appalachian/Caledonide orogen. Nature, London., 222, 124-129.

DEWEY, J.F. (1963) The lower Palaeozoic stratigraphy of central Murrisk, County Mayo, Ireland and the evolution of the south Mayo trough. Quart. J. Geol. Soc. London, 119, 313-344.

DIGGENS, J.N. & ROMANO, M. (1968) The Caradoc rocks around Llyn Colwyd, North Wales. Geol. Journal, 6, 31-48.

DICKINSON, W.R. (1974). Sedimentation within and beside ancient and modern magmatic arcs. In: Dott,Jr., R.H. & Shaver, R.H., (eds)., Modern and ancient geosynclinal sedimentation, Soc. Econ. Paleont. Mineral. Sp. Publ. 19, p.230-239.

DICKINSON, W.R. (1982). Composition of sandstones in circum pacific subduction complexes & forearc basins. Bull of Am. Assoc. Petrol. Geol., 66 , 121-137.

DICKINSON, W.R. & SUCZEK, C.A. (1979). Plate tectonics and sandstone compositions, Am. Assoc. Petrol. Geol., 63, 2164-2182.

DICKINSON, W.R. & VALLONI, R. (1980). Plare settings and provenance of sands in modern ocean basins. Geology., 8, 82-86.

DICKINSON, W.R., BEARD, S.L., BRAKENRIDGE, G.R., ERJAVEC, J.L., FERGUSON, R.C., INMAN, K.F., KNEPP, R.A., LINBERG, F.A. & RYBERG, P.T. (1983). Provenance of North America sandstones in relation to tectonic setting. Bull. Geol. Soc. Am., 94, 222-235.

DIEM, B. (1985). Analytical method for estimating palaeowave climate and water depth from wave ripple marks. Sedimentology, v32 p.705-720.

DIMBERLINE, A.J. & WOODCOCK, N.H. (1987). The southeast margin of the Wenlock Turbidite system, Mid-Wales, Geol. J., 22, 61-72.

DIMROTH, E., IMREH, L., COUSINEAU, P., LEDUC, M., & SANSCHAGRINY, Y. (1985) Palaeogeographic analysis of mafic submarine flows and its use in the exploration for massive sulphide deposits. In: "Evolution of Archean supracrustal sequences" (ed. by L.D. Ayres, P.C. Thurston, K.D. Card, and W. Weber), Geological Association of Canada Special paper 28, p.204-221.

DIXON, J. (1979) The Lower Cretaceous Atkinson Point formation (new name) on the Tuktoyakaktuk peninsula, N.W.T.: A coastal fan delta to marine sequence. Bull. Can. Petrol. Geol., 27, 163-182.

DOTT, R.H. & BOURGEOIS, J. (1982) Hummocky stratification: significance of its variable bedding sequences. Bull. Geol. Soc. Am., 93, 663-680.

DOTT, R.H. & BOURGEOIS, J. (1983) Hummocky stratification: significance of its variable bedding sequences. Reply. Bull. Geol. Soc. Am., 94, 1249-1251.

DUKE, W.L. (1985) Hummocky cross-stratification, tropical hurricanes, and intense winter storms. Sedimentology, 32, 167-194.

DUKE, W.L. (1985). The paleogeography of Paleozic and Mesozoic storm depositional systems: a discussion. J. Geol., 93, p.88-90.

DUKE, W.L. (1987) Reply: Hummocky cross-stratification, tropical hurricanes, and intense winter storms. Sedimentology, 34, 344-359.

DUKE, W.L. (1987). Hummocky cross-stratification, tropical hurricanes, and intense winter storms. Sedimentology, v34, No 2, p.344-359.

DUNKLEY, P.N. (1979). Ordovician volcanicity of the southeast Harlech Dome. In: Harris, A.L., Holland, C.H. & Leake, B.E. (eds). The Caledonides of the British Isles - reviewed. Spec. Publ. geo. Soc. London, 8, 597-601.

DUNNE, L.A. & HEMPTON, M.R. (1984) Deltaic sedimentation in the Lake Hazar pull-apart basin, southeastern Turkey. Sedimentology, 31, 401-412.

DUTTA, P.K. & SUTTNER, L.J. (1986). Alluvial sandstone composition and palaeoclimate, 2: Authigenic mineralogy. J. Sediment. Petrol., 56, 346-358.

EKAS, L.M., INGERSOLL, R.V., BALDRIDGE, W.S. & SHAFIQULLAH, M. (1984). The Chama - El Rito mbr of the Tesuque FM, Espanola Basin, New Mexico. In: New Mexico Geological Society Guidebook. 35th field conference, Rio Grande Rift, northern New Mexico, 1984.

EKDALE, A.A., BROMLEY, R.G. & PEMBERTON, S.G. (1984). Ichnology: The use of trace fossils in sedimentology & sratigraphy. p.157-168.

ELLES, G.L. (1909). The relation of the Ordovician and Silurian rocks of Conway (North Wales). J. Geol. Soc. London., 65, 169-94.

ELLIOTT, T. (1974) Interdistributary bay sequences and their genesis. Sedimentology, 21, 611-622.

ELLIOTT, T. (1986). Deltas: In: Sedimentary Enviroments and Facies. (ed.by H.G.Reading), Blackwell Scientific, Oxford, p.113-154.

ELLIOTT, T. (1986b) Siliclastic shorelines. In: "Sedimentary environments and facies" (ed. by H.G. Reading). Blackwell Scientific, 613 pp.

ETHRIDGE, F.G. & WESCOTT, W.A. (1984) Tectonic setting, recognition, and hydrocarbon reservoir potential of fan delta deposits. In: Sedimentology of gravels and conglomerates (ed. by E.H. Koster and R.J. Steel). Can. Soc. Petrol. Geol. Mem. 10, 217-235.

FARQUHARSON, G.W. (1982) Lacustrine deltas in a Mesozoic alluvial sequence from Camp Hill, Antarctica. Sedimentology, 29, 717-725.

EVANS, A.M. (1979) The East Midlands aulacogen of Caledonian age. Mercian Geol., 7, p.31-42.

EVANS, C.J. & ALLSOP, J.M. (1987) Some geophysical aspects of the deep geology of eastern England. Proc. Yorkshire Geol. Soc., 46,

FEARY, D.A. (1984) The Boambolo formation - a Silurian prograding fan delta sequence in southeastern New South Wales, Australia. Sedimentary Geology, 39, 169-195.

FERGUSON. (1987). Hydraulic and sedimentary controls of channel patterns. In: Rivers, Enviroment process form. (ed.by,K.S.Richards). p.129-158.

FERNANDEZ, H.E. (1969) Notes on the submarine ash flow tuff in Siargo Island, Surigao del Norte. The Phillipine Geologist, 23, 29-36.

FISHER, R.V. (1961). Proposed classification of volcaniclastic sediments and rocks. Geol. Soc. Amer. Bull. 72, 1409-1414.

FISHER, R.V. (1966) Rocks composed of volcanic rock fragments. Earth Sci. Rev., 1, 287-298.

FISHER, R.V. (1971). Features of coarse-grained, high-concentration fluids and their deposits. J. Sed. Petrol. 41, p.916-927.

FISHER, R.V. (1979). Models for pyroclastic surges and pyroclastic flows. J. Volcanol. Geotherm. Res., 6, 305-318.

FISHER, R.V. (1983) Flow transformations in sediment gravity flows. Geology, 11, 273-274.

FISHER, R.V. (1984) Submarine volcaniclastic rocks. In: Marginal Basin Geology: volcanic and associated sedimentary and tectonic processes in modern and ancient marginal basins. (ed. by B.P. Kokelaar and M.F. Howells) Spec. Pub. Geol. Soc. London, 16, p.1-27.

FISKE, R.S. & MATSUDA, T. (1964) Submarine equivelants of ash flows in the Tokiwa Fm., Japan. Am. J. Sci., 262, 76-106.

FISHER, R.V. & SCHMINCKE, H. (1984) Pyroclastic Rocks. Springer-verlag, Berlin. 472 pp.

FISHER, R.V., GLICKEN, H.X. & HOBLITT, R.P. (1987). May 18, 1980. Mount St. Helens deposits in south Coldwater creek, Washington. J. Geophys. Res, 92, p. 10267-10283.

FITCH, F.J. (1967) Ignimbrite volcanism in North Wales. Bull. volcan., 30, 199-219.

FITTON, J.G. & HUGHES, D.J. (1970) Volcanism and plate tectonics in the British Ordovician. Earth Planet. Sci. Letters., 8, 223-228.

FLINT, S., CLEMMEY, H. & TURNER, P. (1986) The Lower Cretaceous Way Group of Northern Chile: An alluvial fan-fan delta complex. Sedimentary Geology, 46, 1-22.

FLORES, R.M. (1975) Short headed stream delta: model for Pennsylvanian Haymond Formation. Bull. Am. Assoc. Petrol. Geol., 59, 2288-2301.

FLORES, R.M. & FERM, J.C.(1970) A recent model for Pennsylvanian deposition in the Marathon Basin, West Texas. J. Sediment. Petrol., 40, 621-628.

FORTEY, R.A. (1984) Global earlier Ordovician transgressions and regressions and their biological implications. In: "Aspects of the Ordovician System" (ed. by D.L. Bruton). Palaeontological Contributions from the University of Oslo no. 295, Universitetsforlaget, p. 37-50.

FRANCIS, E.H. & HOWELLS, M.F. (1973). Transgressive welded ash-flow tuffs among the Ordovician sediments of NW Snowdonia, N. Wales. J. geol. Soc. Lond., 129, 621-41.

FRANZINELLI, E. & POTTER, P.E. (1983). Petrology, chemistry and texture of modern river sands, Amazon river system : J. Geol. 91, 23-39.

FREUNDT, A. & SCHMINCKE, H.U. (1986). Emplacement of small volume pyroclastic flows at Laacher See (East Eifel, Germany). Bull. Volcanol., 98, 39-59.

FREUNDT, A. & SCHMINCKE, H.U. (1985). Lithic enriched segregation bodies in pyroclastic flows deposits of Laacher see volcano (E. Eifel, Germany). J. volcanol. Geotherm. Res., 25, 193-224.

FREY, R.W. & PEMBERTON, S.G. (1984). Trace fossils facies models. In: Facies Models. (ed. by R.G.Walker) Geoscience Canada reprint series 1. p.189-208.

FRIEDMAN, I, LONG, W. & SMITH, R.L. (1963) Viscosity and water content of

rhyolite glass. J. geophys. Res., 68, 6523-6535.

FRIEND, P.F. (1983) Towards a field classification of alluvial architecture and sequence. In: Modern and Ancient fluvial systems (ed. by J.D. Collinson and J. Lewin). Spec. Pub. Int. Assoc. Sediment., 6, 345-354.

FRIEND, P.F., SLATER, M.J. & WILLIAMS, R.C. (1979). Vertical and lateral building of sandstone bodies in the Ebro river. Geol. Soc. London., 136, 39-46.

FRITZ, W.J. & HARRISON, W.J. (1985) Early Tertiary volcaniclastic deposits of the Northern Rocky Mountains. In. Paleogeography of West Central United States. (ed. by R.M. Flores and S.S. Kaplan), 383-402.

FROSTICK, L.E., RENAULT, R.W., REID, I, & TIERCELIN, J.J. (1986) (EDS.) Sedimentation in the African Rifts. Geological Society Special Publication no. 25, 382pp.

FURSICH, F.T. (1982) Rhythmic bedding and shell bed formation in the Upper Jurassic of East Greenland. In: Cyclic and event stratification (ed. by G. Einsele and A Seilacher), Springer-Verlag, Berlin, p.208-222.

GALLOWAY, W.E. (1975) Process framework for describing the morphological and stratigraphic evolution of deltaic depositional systems. In: Deltas, models for exploration. (ed. by M.L. Broussard). Houston Geol. Soc., 87-98.

GALLOWAY, W.E. (1976) Sediments and stratigraphic framework of the Copper River fan-delta, Alaska. J. Sediment. Petrol., 46, 726-737.

GALLOWAY, W.E. & HOBDAY, D.K. (1983) Terrigenous clastic depositional systems: applications to Petroleum, coal and uranium exploration. Springer-Verlag, New York. 423pp.

GALLOWAY, W.E.(1981) Depositional architecture of Cenozoic Gulf Coastal plain fluvial systems. In: Recent and Ancient Nonmarine depositional environments; models for exploration. (ed. by F.G. Ethridge and R.M. Flores). Spec. Pub. Soc. Econ. Paleont. Min., 31, 127-155.

GARRELS, R.M. & Mackenzie, F.T. (1971). Evolution of sedimentary rocks. W.W. Norton & Co., Inc., New York, p.1-397.

GARZANTI, E. (1986) Source rock versus sedimentary control on the mineralogy of deltaic volcanicarenites (Upper Triassic, Northern Italy) J. Sediment. Petrol., 56, 267-275.

GEORGE, T.N. (1961) North Wales. British Regional Geology. London.

GLOPPEN, T.G. & STEEL, R.J. (1981) The deposits, internal structure and geometry in six alluvial fan-fan delta bodies (Devonian, Norway) - a study in the significance of bedding sequences. In: Recent and Ancient Nonmarine depositional environments: models for exploration (ed. by F.G. Ethridge and R.M. Flores). Spec. Pub. Soc. Econ. Paleont. Min., 31, 49-69.

GOLDRING, R. & AIGNER, Y. (1982) Scour and fill: the significance of event separation. In: Cyclic and event stratification (ed. by G. Einsele and A Seilacher), Springer-Verlag, Berlin,1 p.354-362.

GOLDRING, R. & BRIDGES, P. (1973) Sublittoral sheet sandstones. J.

Sediment. Petrol., 43, 736-747.

GOLOMBEK, M.P. & CARR, M.J. (1978). Tidal triggering of seismic avolcanic phenomena during the 1879-1880 eruption of Islas Quemadas volcano in El. Salvador, Central America. Journal Volcanol Geotherm Res., 3, 299-307.

GREENWOOD, B. & SHERMAN, D.J. (1986). Hummocky cross-stratification in the surf zone: flow parameters and bedding genesis. Sedimentology, 33, p.33-45.

HAGDORN, H. (1985). Immigration of crinoids into the German Muschelkalk basin. In: U. Bayer & A. Sulacher (eds). Sedimentary & Evolutionary Cycles: Springer-Verlag, Berlin, p.237-254.

HAMBLIN, A.P. & WALKER, R.G. (1979). Storm-dominated shallow marine deposits: the Fernie-Kootenay (Jurassic) transition, southern Rocky Mountains. Can. J. Earth Sci., 16, p. 1673-1690.

HAMILL, P.F. & BALLANCE, P.F. (1985) Heavy mineral rich beach sands of the Waitakere coast, Auckland, New Zealand. N.Z. J. Geol. Geophy., 28, 503-512.

HAMPTON, M.A. (1972) The role of subaqueous debris flows in generating turbidity currents. J. Sediment. Petrol., 42, 775-793.

HARBAUGH, D.W. & DICKINSON, W.R. (1981) Depositional facies of Mississippian clastics, Antler foreland basin, Central Diamond Mountains, Nevada. J. Sediment. Petrol., 51, 1223-1234.

HARKER, A. (1889) The Bala volcanic series of Caernarvonshire. Sedgewick Prize Essay, Cambridge.

HARLAND, W.B. & GAYER, R.A. (1972) The Arctic Caledonides and earlier oceans. Geological Magazine, 109, 289-384.

HARMS, J.C. , SOUTHARD, J.B., SPEARING, D.R. & WALKER, R.G. (1975). Depositional enviroments as interpreted from primary sedimentary structures and stratification sequences. Soc. econ. Paleont. Miner., Short Course No 2, 161 pp.

HARMS, J.C., SOUTHARD, J.B. & WALKER, R.G. (1982). Structures and Sequences in Clastic Rocks. Soc. econ. Paleont. Miner., Short Course No 9.

HARVEY, A.M. (1984) Debris flows and fluvial deposits in Spanish Quaternary alluvial fans: implications for fan morphology. In: Sedimentology of gravels and conglomerates (ed. by E.H. Koster and R.J. Steel) Can. Soc. Petrol. Geol. Mem 10, 123-132.

HASZELDINE, R.S. (1983). Descending tabular XB sets and bounding surfaces from a fluvial channel in the upper carboniferous coalfield of north-east New England. In: Modern and Ancient fluvial systems (ed. by J.D. Collinson and J. Lewin). Spec. Pub. Int. Assoc. Sediment., **6**, p.449-456.

HAYES, B.J.R. & SMITH, D.G. (1987) Discussion of: Cardium Formation 6: Stratigraphic framework for the Cardium in subsurface. Bull. Can. Petrol. Geol., 35, 363-365.

HAYES, M.O. & MICHEL, J. (1982) Shoreline sedimentation within a forearc embayment, Lower Cook Inlet, Alaska. J. Sediment. Petrol., 52, 251-263.

HAYWARD, A.B. (1985) Coastal alluvial fans and associated marine facies in the Miocene of S.W. Turkey. In: Modern and Ancient fluvial systems (ed. by J.D. Collinson and J. Lewin). Spec. Pub. Int. Assoc. Sediment., 6, 323-336.

HEIN, F.J. (1987) Deep sea and fluvial braided channel conglomerates: a comparison of two case studies. In: Sedimentology of gravels and conglomerates. (ed. by E.H. Koster and R.J. Steel) Can. Soc. Petrol. Geol. Mem. 10, p. 33-50.

HENRY & PRICE, (1984). Variations in caldera development in the tertiary volcanic field of Trans Pecos, Texas. J.Geophys. Res., 89, 8765-8786.

HEWARD, A.P. (1981) A review of wave dominated clastic shoreline deposits. Earth Science Reviews, 17, 223-276.

HILDRETH, W. (1981). Gradients in silicic magma chambers. Implications for lithospheric magmatism. JGR, 36, no 10, 10153-10192.

HINE, A.C. & BOOTHROYD, J.C. (1978) Morphology, processes and sedimentary history of a glacial outwash plain shoreline, Southern Iceland. J. Sediment. Petrol., 48, 901-920.

HOBDAY, D.K. (1974) Interactions between fluvial and marine processes in the lower part of the Late Precambrian Vadso Group, Finnmark. Norges Geologiske undersokelse, 303, 39-56.

HOBDAY, D.K. & READING, H.G. (1972) Fair weather versus storm processes in shallow marine sand bar sequences in the late Pre-Cambrian of Finnmark, North Norway. J. Sediment. Petrol., 42, 318-324.

HOBLITT, R.P., MILLER, C.D. & VALLANCE, J.W. (1981). Origin and stratigraphy of the deposit produced by the May 18 directed blast. In: Lipman, P.W. & Mullineaux, D.R. (eds). The 1980 eruptions of Mount St.Helens, Washington. U.S.A. Geol. Survey Prof. Paper 1250, p. 401-419.

HOLLAND, C.H. (1959) 1. The Ludlovian and Downtonian rocks of the Knighton District, Radnorshire. Quart. J. Geol. Soc. London., 114, 449-482.

HOLMES, A. (1965) Principles of Physical Geology. 2nd edition. New York. The Roland Press Co., 1288 pp.

HOOKE, R. Le. B. (1967) Processes on arid region alluvial fans. J. Geology, 75, 438-460.

HOOKE, R. Le. B. (1968) Steady state relationships on arid region alluvial fans. J. Geology, 75, 438-460.

HOWARD, J.D. & REINECK, H.E. (1981) Depositional facies of high energy beach to offshore sequence: Comparison with low energy sequence. Bull. Am. Assoc. Petrol. Geol., 65, 807-830.

HOWELLS, M.F. & LEVERIDGE, B.E. (1980) The Capel Curig Volcanic Formation. Rept. Inst. Geol. Sci., 80/6, 1-23.

HOWELLS, M.F., CAMPBELL, S.D.G., & REEDMAN, A.J. (1985) Isolated pods of subaqueous welded ash flow tuffs: a distal facies of the Capel Curig Volcanic Formation (Ordovician), North Wales. Geol. Magazine, 122, 175-180.

HOWELLS, M.F., CAMPBELL, S.D.G. & REEDMAN, A.?. (1986) Isolated pods of subaqueous welded ash flow tuff: a distal facies of the Capel Curig Volcanic Formation (Ordovician), North Wales. Geol. Mag., 123, p.590.

HOWELLS, M.F., FRANCIS, E.H., LEVERIDGE, B.E. & EVANS, C.D.R. (1978) Capel Curig and Betws-y-Coed: Description of 1:25,000 sheet SH75. Classical areas of British Geology, Institute of Geological Sciences, H.M.S.O., London, 73 pp.

HOWELLS, M.F., LEVERIDGE, B.E., & EVANS, C.D.R. (1973) Ordovician ash flow tuffs in eastern Snowdonia. Rept. Inst. Geol. Sci., 83/1, 11-15.

HOWELLS, M.F., LEVERIDGE, B.E., ADDISON, R., EVANS, C.D.R., & NUTT, M.C.J. (1980)The Capel Curig Volcanic Formation, Snowdonia, North Wales: variations in ash-flow tuffs related to emplacement environment. In: gslsp8, 611-618.

HOWELLS, M.F., LEVERIDGE, B.E., & REEDMAN, A.J. (1981b) Snowdonia. Rocks and Fossils series, No.1, Unwin Paperbacks, London, 119 pp.

HOWELLS, M.F., LEVERIDGE, B.E., ADDISON, R. & REEDMAN, A.J. (1983) The lithostratigraphic subdivision of the Ordovician underlying the Snowdon and Crafnant volcanic groups, North Wales. Rept. Inst. Geol. Sci., 83/1, 11-15.

HOWELLS, M.F., LEVERIDGE, B.E., EVANS, C.D.R., & NUTT, M.J.C. (1981a) Dolgarrog: Description of 1:25,000 Geological sheet SH76. Classical areas of British Geology, Institute of Geological Sciences, H.M.S.O., London, 89 pp.

HOWELLS, M.F., LEVERIDGE, B.E., ADDISON, R., EVANS, C.D.R. and NUTT, M.J.C. (1979) The Capel Curig volcanic formation, Snowdonia, North Wales; variations in ash flow tuffs related to emplacement environment. In: "The Caledonides of the British Isles. Geol. Soc. London., Special Publication, p. 611-618.

HOWELLS, M.F., LEVERIDGE, B.F., ADDISON, R. & REEDMAN, A.J. (1983) The lithostratigraphic subdivision of the Ordovician underlying the Snowdon and Crafnant volcanic groups, North Wales. Report. Inst. Geol. Sci., no. 83/1, 11-15.

HUFF, W.D. (1983). Correlation of middle ordovician K. bentonites based on chemical fingerprinting. J.Geol., 91, 657-669.

HUNTER, R.E. & CLIFTON (1982) Cyclic deposits and hummocky cross-stratification of probable storm origin in Upper Cretaceous rocks of the Cape Sebawtian area, SW Oregon. J. Sediment. Petrol., 52, 127-143.

HUNTER, R.E., CLIFTON, H.E., & PHILLIPS, R.L. (1979). Depositional processes, sedimentary structures, and predicted vent sequence in barred nearshore systems, southern Oregon coast. J. Sediment. Petrol., 49, 711-726.

HUNTER, R.E., CLIFTON, H.E. & PHILLIPS, R.L. (1979). Depositional

processes, sedimentary structures, and predicted vertical sequence in barred nearshore systems, Southern Oregon coast. J. Sediment. Petrol., 49, 711-726.

HUPPERT, H.E., TURNER, J.S., CAREY, S.N., SPARKS, R.S.J. & HALLWORTH, M.A. (1986) A laboratory simulation of pyroclastic flows downslopes. J. Volcanol. Geotherm. Res., 30, 179-199.

HURST, J.M. (1976). The environment of deposition of the Caradocalternata limestone & contiguos deposits of salop. Geol. J., 14, 15-40.

HURST, J.M. (1979). The stratigraphy & brachiopods of the upper part of the type Caradoc of south Salop. Bull. Br. Mus. Nat. Hist., 32, 183-304.

INGERSOLL, R.V. & SUCZEK, C.A. (1979) Petrology and provenance of neogene sand from Nicobar and Bengal Fan, DSDP sites 211 & 218. J. Sediment. Petrol., 49, 1217-1228.

INGERSOLL, R.V., BULLARD, T.F., FORD, R.L., GRIMM, J.P., PICKLE, J.D. & SARES, S.W. (1984) The effect of grain size on detrital modes: a test of the Gazzi-Dickinson point counting method. J. Sediment. Petrol., 54, 103-116.

JAANUSSON, V. (1984) What is so special about the Ordovician? In: "Aspects of the Ordovician System 1-3. Palaeontological Contributions from the University of Oslo No. 295., Universitetsforlaget (ed. by. D. L. Bruton)., p. 1-3.

JACKSON, R.G. (1976). Depositional model of point bars in the lower Wabash river. J. Sediment. Petrol., 46, 579-594.

JACKSON, R.G. (1978) Preliminary evaluation of lithofacies models for meandering alluvial systems. In: "Fluvial sedimentology" (Ed.. by A.D. Miall). Can. Soc. Petrol. Geol. Mem. 5, 543-576.

JAMES, D.M.D. (1973). The Garnedd-wen Formation (Ashgillian) of the Towyn-Abergynolwyn district, Merionethshire. Geol. Mag., 110, 145-52.

JAMES, D.M.D. (1975). Caradoc turbidites at Poppit Sands (Pembrokeshire), Wales. Geol. Mag., 112, 295-304.

JAMES, D.M.D. (1983a). Sedimentation of deep-water slope-base and inner-fan deposits - the Drosgol Formation (Ashgill), West Central Wales. Sedim. Geol., 34, 21-40.

JOHNSON, H.D. (1977) Shallow marine sand bar sequences: an example from the late Precambrian of North Norway. Sedimentology, 24, 245-270.

JOHNSON, H.D. & BALDWIN, C.T. (1986) Shallow siliciclastic seas. In: Sedimentary Environments and Facies (ed. by H.G. Reading),p. 229-282, Blackwell Scientific Publications, 615 pp.

JONES, O.T. (1938) On the evolution of a geosyncline. Quart. J. Geol. Soc. London, 94, lx-cx.

JONES, O.T. (1940) The geology of the Colwyn Bay district; a study of submarine slumping during the Salopian period. Quart. J. Geol. Soc. London, 95, 335-382.

JONES, O.T. (1956) The geological evolution of Wales and the adjacent

regions. Quart. J. geol. Soc. London, 111, 323-351.

JONES, O.T.& PUGH, W.J. (1949) An early Ordovician shoreline in Radnorshire, near Builth Wells. Quart. J. geol. Soc. London, 105, 65-99.

JOPLING, A.V. (1965) Hydraulic factors controlling the shape of laminae in laboratory deltas. J. Sediment. Petrol., 35, 777-791.

KELLING, G. (1964). The turbidite concept in Britain. In: Turbidites: Devs Sedimentol. no.3, (ed. by A.H. Bouma & A. Brouwer), p. 75-92.

KINGSLEY, C.S. (1984) Dagbreek fan delta: An alluvial placer to prodelta sequence in the Proterozoic Welkom Goldfield, Witwatersrand, South Africa. In: Sedimentology of gravels and conglomerates. (ed. by E.H. Koster and R.J. Steel) Can. Soc. Petrol. Geol. Mem. 10, 321-330.

KIRSCHVINK, J.L. (1984) Ultrafine-grained magnetite in deep-sea sediments: possible bacterial magnetofossils. Geology., 12, 559-562.

KIRSCHVINK, J.L. & CHANG, S.B.R. (1984) Ultrafine-grained magnetite in deep sea sediments: possible bacterial magnetofossils. Geology, 12, 559-562.

KLEIN, G. de V & RYDER, T.A. (1978). Tidal circulation patterns in PE, paleozoic & gretaceous epeiric & miocleinal shelf seas. Bull. Geol. Soc. Am., 89, 1050-1058.

KLEIN, G. de V. (1974) Estimates of water depths from analysis of barrier islands and deltaic sedimentary sequences. Geology, 2, 409-412.

KLEIN, G. de V. & MARSAGLIA, K.M. (1987) Hummocky cross-stratification, tropical storms and intense winter storms: A discussion. Sedimentology, 34, 333-337.

KLEINSPEHN, K.L., STEEL, R.J., JOHANNESSEN, E. & NETLAND, A. (1984) Conglomeratic fan-delta sequences, Late Carboniferous - Early Permian, Western Spitsbergern. In: Sedimentology of gravels and conglomerates. (ed. by E.H. Koster and R.J. Steel) Can. Soc. Petrol. Geol. Mem. 10, 279-294.

KOKELAAR, B.P. (1979). Tremadoc to Llanvrin volcanism on the southeast side of the Harlech Dome (Rhobell Fawr), North Wales. In: Harris, A.L., Holland, C.H. & Leake,B.E. (eds). The Caledonides of the British Isles-reviewed. Spec. Publ. geol. Soc. London., 8, 591-6.

KOKELAAR, B.P. (1982) The fluidization of wet sediments during the emplacement and cooling of various igneous bodies. J. geol. Soc. London., 139, 21-33.

KOKELAAR, B.P., BEVINS, R.E., & ROACH, R.A. (1985) Submarine silicic volcanism and associated sedimentary and tectonic processes, Ramsay Island, SW Wales. J . Geol. Soc. London, 142, 591-613.

KOKELAAR, B.P., HOWELLS, M.F., BEVINS, R.E., ROACH, R.A. & DUNKLEY, P.N. (1984) The Ordovician marginal basin of Wales. In: Marginal Basin Geology: volcanic and associated sedimentary and tectonic processes in modern and ancient marginal basins. (ed. by B.P. Kokelaar and M.F. Howells) Spec. Pub. Geol. Soc. London, 16, 245-270.

KOKELAAR, B.P., HOWELLS, M.F., BEVINS, R.E., and ROACH, R.A. (1984) Volcanic and associated sedimentary and tectonic processes in the Ordovician marginal basin of Wales: a field guide. IN: "Marginal Basin geology: volcanic and associated sedimentary and tectonic processes in modern and ancient marginal basins. (ed. by B.P. Kokelaar and M.F. Howells), p. 291-322.

KOKELAAR, P. (1986) Magma water interactions in subaqueous and emergent basaltic volcanism. Bull. volc., 48, 275-289.

KOMAR, P.D. (1971) Hydraulic jumps in turbidity currents. Bull. Geol. Soc. Am., 82, 1477-1488.

KOMAR, P.D. (1983) Shapes of streamlined islands on earth and mars: experiments and analysis of the minimum drag force. Geology, 11, 651-654.

KOSTACHUK, R.A., MacDONALD, G.M. & PUTNAM, P.E. (1986) Depositional process and alluvial fan drainage basin morphometric relationships near Banff, Alberta, Canada. Earth Surface Processes, 11, 471-484.

KREISA, R.D. & BAMBACH, R.K. (1982) The role of storm processes in generating shell beds in Palaeozoic shelf environments. In: Cyclic and event stratification (ed. by G. Einsele and A Seilacher), Springer-Verlag, Berlin, p.200-207.

KREISA, R.D. & MOIOLA, R.J. (1986) Sigmoidal tidal bundles and other tide-generated sedimentary structures of te Curtis Formation, Utah. Bull. Geol. Soc. Am., 97, 381-387.

KUENEN, P.H. (1853). Graded bedding, with observations on Lower Paleozoic rocks of Britain. K. Nederl Akad. Westens., Afd. Natuurk, 20, 1-47.

KUENEN, P.H. & MIGLIORINI , F. (1950) Turbidity currents as a cause of graded bedding. J. Geol., 58, 91-127.

KUENZI, W.D., HORST, O.H. & McGHEE, R.V. (1979) Effect of volcanic activity on fluvio-deltaic sedimentation in a modern arc-trench gap, SW Guatemala. \Bull. Geol. Soc. Am., 90, 827-838.

LAMONT, A. (1946). B.B. Bancrofts geological work. 1. Cardovician lavas and sediments in the Capel Curig district, Caernarvonshire. Quarry Managers Journal, xx, 235-239.

LAPWORTH, C. (1879) On the tripartite classification of the Lower Palaoeozoic rocks. Geol. Mag., 16, 1-15.

LARSEN, V. & STEEL, R.J. (1978) The sedimentary history of a debris flow dominated alluvial fan - study of textural inversion. Sedimentology, 25, 37-59.

LAWSON, D.E. (1982). Mobilization movement and deposition of active subaeriel sediment flows, Matanuska Glacier, Alaska. J.Geol., 90, 279-300.

Le BAS, M.J. (1972) Caledonian igneous rocks beneath central and eastern England. Proc. Yorkshire Geol. Soc., 39, 71-86.

LEAT, P.T., JACKSON, S.E., THORPE, S.E., & STILLMAN, C.J. (1986) Geochemistry of bimodal basalt subalkaline-peralkaline rhyolite

provinces within the Southern British Caledonides. J. Geol. Soc. London, 143, 259-273.

LECKIE, D.A. & WALKER, R.G. (1982) Storm and tide dominated shorelines in Cretceous Moosebar-Lower Gates interval - outcrop equivelants of deep basin gas trap in Western Canada. Bull. Am. Assoc. Petrol. Geol., 66, 138-157.

LEITHOLD, E.H. & BOURGEOIS, J. (1984) Characteristics of coarse-grained sequences deposited in nearshore wave dominated environments-examples from the Miocene of SW Oregon. Sedimentology, 25, 37-59.

LEEDER, M.R. (1983). On the interactions between rurbulent flow, sediment transport and bedform mechanics in channelised flows. IN: Modern and Ancient fluvial systems (ed. by J.D. Collinson and J. Lewin). Spec. Pub. Int. Assoc. Sediment., 6, 5-18.

LEEDER, M.R. & GAWTHORPE, R.L. (1987) Sedimentary models for extensional tilt-block/half-graben basins. In: "Continental extensional tectonics" (ed. by M.F. Coward, J.F. Dewey, & P.L. Hancock). Geological society special publication no. 28, p. 139-152.

LEVELL, B.K. (1980) A late Precambrian tidal shelf deposit, the Lower Sandfjord Formation, Finmark, North Norway. Sedimentology, 27, 539-557.

LINDSTROM, M. (1974). Volcanic contribution to Ordovician pelagic sediments. J. Sed. Petrol. 44, 287-291.

LINK, M.W. (1984). Fluvial facies of the Miocene ridge route formation, ridge basin, California. Sed.Geol., 38, 263-285.

LIPMAN, P.W. (1984) The roots of ash flow calderas in western North America: Windows into the tops of Batholiths. J. Geophys. Research, 89, 8801-8841.

LIPMAN, P.W., MOORE, J.G. & SWANSON, D.A. (1981). The bulging of the north flank before the Mat 18 eruption, geodetic data.

LOCKLEY, M.G. (1980a) The Caradoc faunal associations of the area between Bala and Dinas Mawddwy, North Wales. Bull. Br. Mus. nat. Hist. (Geol.), 33, 165-235.

LOCKLEY, M.G. (1980b) The geology of the Llanuwchllyn to Llanymawddwy area, south Gwynedd, North Wales. Geol. J., 15, 21-41.

LOCKLEY, M.G. (1983). A review of brachiopod dominated palaeocommunities from the type Ordovician. Palaeontology, 26, 111-145.

LOCKLEY, M.G. (1984). Faunas in a volcaniclastic debris flow from the Welsh basin. A synthesis of paleological and volcanological observations. In: "Aspects of the Ordovician System" (ed. by D.L. Bruton). Palaeontological contr. from University of Oslo no295. Universitetsforlaget., p.195-201.

LOWE, D.R. (1979). Sediment gravity flows: their classification and some problems of application to natural flows and deposits. Spec. pub. SEPM v27, p.75-82.

LOWE, D.R. (1982). Sediment gravity flows: Depositional models with special

reference to the deposits of high density turbidity currents. J. Sediment. Petrol., 52, 279-297.

LOWMAN, R.D.W. & BLOXAM, T.W. (1981) The petrology of the Lower Palaeozoic Fishguard Volcanic Group & associated rocks east of Fishguard, North Pembrokeshire (Dyfed), South Wales. J. Geol. Soc. London, 138, 47-68.

MACK, G.H. (1982). Sed photo 2m sheetfold flood waters emerge fanhead 5pt 25angle. J. Sediment. Petrol., 52, 250.

MACK, G.H. (1984) Exceptions to the realtionship between plate-tectonics and sandstone composition. J. Sediment. Petrol., 54, 212-220.

MARSAGLIA, K.M. & KLEIN, G de V. (1983) The palaeogeography of palaeozoic and mesozoic storm depositional systems. J. Geol., 9, 117-142.

MACKIE, S. & SMALLWOOD, S. (1987). A revised stratigraphy and sedimentology of the Abergwesyn-Pumpsaint area, Mid-Wales. Geol. J., 22, 45-60.

MASSARI, F. (1984) Resedimented conglomerates of a Miocene fan delta complex, Southern Alps, Italy. In: Sedimentology of gravels and conglomerates. (Ed. by E.H. Koster and R.J. Steel) Can. Soc. Petrol. Geol. Mem. 10, 259-278.

MATHISEN, M.E. (1986). Diogenesis of plio pleistocene non marine sandstones, Cagoyan Basin, Phillipines : early development of secondry porosity in volcanic sandstones. In: AAPG Mem 30, p.177-193.

MATHISEN,M.E. & VONDRA, C.F. (1983) The fluvial and pyroclastic deposits of the Cagayan Basin, Northern Luzon, Philippines - an example of non-marine volcaniclastic sedimentation in an interarc basin. Sedimentology, 30, 369-392.

MAYNARD, J.B., VALLONI, R. & YU, H. (1982). Composition of modern deep sea sands from arc related basins. In: Trench - Forearc Geology (ed. by J.K. Legget). Special publ. Geol. Soc. London. v10 p.551-561.

McBIRNEY, A.R. (1968). Second additional theory on the origin of fiamme in ignimbrites. Nature, 217, 938.

McGOWEN, J.H. (1971) Gum Hollow fan delta, Nueces Bay, Texas. Univ. Texas Bureau of Economic Geology, Rept. Inv., 69, 91 pp.

McGOWEN, J.H. & SCOTT. A.J. (1974) Fan delta deposition: processes, facies and stratigraphic analogues (Abstract). Am. Assoc. Petrol. Geol/Soc. Econ. Palaeont. Min., Abstracts, 60-61

McKENZIE, D.P. (1978) Some remarks on the development of sedimentary basins. Earth Planet. Sci letters.,v40,25-32.

McPHERSON, J.G., SHANMUGAM, G. & MOIOLA, R.J. (1987). Fan deltas and braid deltas: varietiesof course grained deltas. Bull. Geol. Soc. Am., 99, 331-340.

McPHERSON, J.G., SHANMUGAM, G. & MOIOLA, R.J. (1988). Fan deltas and braid deltas: conceptual problems. In: Fan Deltas & Related systems, Sedimentology and Tectonic settings. (ed. by. W. Nemec & R.J. Steel). Blackie, Glasgow. (in press)

MIALL, A.D. (1976) Facies models 4: Deltas. Geoscience Canada, 3, p 215-227.

MIALL, A.D. (1977) A review of the Braided river depositional environment. Earth Science Reviews, 13, 1-16.

MIALL, A.D. (1984) Variations in fluvial style in the Lower Cenozoic synorogenic sediments of the Canadian Arctic Islands. Sedimentary Geology., 38, 499-523.

MIALL, A.D. (1985). Architectural element analysis: A new method of facies analysis applied to fluvial deposits. Earth Science Reviews., 22, 261-308.

MIDDLETON, G.V. & SOUTHARD, J.B. (1978) Mechanics of sediment movement. SPEM Short Course no. 3.

MILLS, A.A. (1984) Pillow lavas and the Leidonfrost effect. J. Geol. Soc. London., 141, 183-186.

MITCHELL, A.H. & READING, H.G. (1971) Evolution of Island arcs. J. geology., 79, 253-284.

MITCHELL, A.H.G. (1970) Facies of an Early Miocene volcanic arc, Malekula Island, New Hebrides. Sedimentology, 14, 201-243.

MITCHELL, A.H.G., HERNANDEZ, F. & DELA CRUZ, A.P. (1986) Cenozoic evolution of the Philippine Archipelago. J. Southeast Asian Earth Science., 1, 3-22.

MITCHELL, G.H. (1957) Ordovician volcanoes. Advanc. Sci., 14, 34-47.

MITCHUM, R.M., VAIL, P.R. & THOMPSON, S. (1977). Seismic strarigraphy and global changes of sea level, part two: the depositional sequence as a basic unit for stratigraphic analysis. Am. Assoc. Petrol. Geol. Mem., 26, 979-988.

MOORE, G.T. & ASQUITH, D.O. (1971) Delta: term and concept. Geol. Soc. Am. Bull., 82, 2563-2568.

MOORE, J.G. & ALBEE, W.C. (1981). Topographic and strucyural changes. March-July 1980 photogrammetric data.

MOORE, J.G. & SISSON, T.W. (1981). Deposits and effects of the May 18 pyroclastic surge. In: Lipman, P.W. & Mullineaux, D.R. (eds). The 1980 eruptions of Mount St. Helens, Washington. U.S.A. Geol. Survey Prof. Paper 1250, p.421-438.

MORAD, S. & ALDAHAN, A.A. (1986). Alleration of detrital Fe. T oxides in sedimentary rocks. Bull. Geol. Soc. Am., 97 , 567-578.

MUGGE, O. (1893) Untersuchungen uber die Lenneporphyre in Westfalen und den angrenzenden Gebieten. Neues Jahrb. Beilage Band, 8, 713-714.

MURCHISON, R.I (1839). The Silurian System founded on geological researches in the counties of Salop, Hereford, Radnor, Montgomery, Caermarthen, Brecon, Pembroke, Monmouth, Gloucester, Worcester, and Stafford; with descriptions of the coal-fields and overlying formations.

NEEDHAM, R.S. (1978) Giant-scale hydroplastic deformation structures formed

by the loading of basalt onto water saturated sand, Middle Proterozoic, Northern Territory, Australia. Sed., 25, 285-295.

SMITH, A.G., HURLEY, A.M. & BRIDEN, J.C. (1980) Phanerozoic palaeocontinental world maps. Cambridge University Press, 98pp.

NEMEC, W. & MUSZYNSKI, A.(1982) Volcaniclastic alluvial aprons in the Tertiary of Sofia district (Bulgaria). Annales Societatis Geologorum Polonaie. Rocznik Polskiego Towarzystwa geologicznego, 52, 239-303.

NEMEC, W. & STEEL, R.J. (1984) Alluvial and coastal conglomerates: their significance and some comments on gravelly mass flow deposits. In: Sedimentology of gravels and conglomerates. (ed. by E.H. Koster and R.J. Steel) Can. Soc. Petrol. Geol. Mem. 10, 1-33.

NEMEC, W. STEEL, R.J., POREBSKI, S.J. & SPINNANGR, A. (1984) Domba conglomerate, Devonian, Norway: process and lateral variability in a mass flow dominated lacustrine fan delta. In: Sedimentology of gravels and conglomerates. (ed. by E.H. Koster and R.J. Steel) Can. Soc. Petrol. Geol. Mem. 10, 295-320.

NILSEN, T.H. (1982). Alluvial fan deposits in Schole, P.A. & Spearing, D. eds, Sandstone Depositional Enviroments. Bull. Am. Assoc. Petrol. Geol., memoir 31.

NOTTVERDT, A. & KREISA, R.D. (1987). Model for the combined-flow origin of hummocky cross-stratification. Geology, 15, 357-361.

OJEDA, H.A.O (1982) Structural framework, stratigraphy and evolution of Brazilian Marginal basins, Brazil. Bull. Am. Assoc. Petrol. Geol., 66, 732-749.

OKADA, H. (1971). Classification of sandstone: analysis and proposal. J. Geol., 79, 509-525.

OKOLO, S.A. (1983). Fluvial distributary channels in the Fletcher Bank Grit (Namurion R2b), at Ramsbottom, Lancashire, England. IN: Modern and Ancient fluvial systems (ed. by J.D. Collinson and J. Lewin). Spec. Pub. Int. Assoc. Sediment., 6,, 421-433.

OLIVER, R.L. (1954) Welded tuffs in the Borrowdale volcanic series, English Lake district, with a note on similar rocks in Wales. Geological Mag., 91, 473-483.

ORTON, G.J. (1983). Canyon fjord formation: petrography and grain size characteristics unpublished. BSc thesis University of Toronto.

ORTON, G.J. (1986) Tectonic control of mid-Ordovician volcaniclastic sedimentation , North Wales. Abstract. Bull. Am. Assoc. Petrol. Geol., 70, 628.

ORTON, G.J. (1988). A spectrum of mid-ordovician fan deltas & braidplain deltas, North Wales: a consequence of varying fluvial clastic input. In: Fan Deltas & Related systems, Sedimentology and Tectonic settings. (ed. by. W. Nemec & R.J. Steel). Blackie, Glasgow.

PADGHAM, W.A. (1980) An Archean ignimbrite at Yellowknife and its relationship to the Kam Formation basalts. Precambrian Research, 12, 99-113.

PEARCE, J.A. & CANN, J.R. (1973). Tectonic setting of basic volcanic rocks determined using trace element analyses. Earth planet. Sci. Lett., 19, 290-300.

PETTERS, S.W. (1977) Stratigraphic evolution of the Benue trough and itsimplications for the Upper Cretaceous of West Africa. J. Geol., 86, p.311-322.

PETTIJOHN, F.J. (1950) Turbidites and greywackes - A discussion. J. Geol., 58, p.169.

PETTIJOHN, F.J., POTTER, D.E. & SIEVER, R. (1973). Sand and sandstone. Springer verlag Berlin 618 pp.

PHARAOH, T.C., MERRIMAN, R.J., WEBB, P.C. & BECKINSALE, R.D. (1988) The concealed Caledonides of eastern England: preliminary results of a multi-disciplinary study. Proc. Yorkshire Geol. Soc., 46, 325-369.

PHILLIPS, W.E.A., STILLMAN, C.J. & MURPHY, T. (1976) A Caledonian plate-tectonic model. J. Geol. Soc. London., 132, 579-609.

PICKERILL, R.K. & BRENCHLEY, P.J. (1979). Caradoc marine benthic communities of the south Berwyn Hills, North Wales. Palaeontology, 22, 229-64.

PIERSON, T.C. (1981). Dominant particles support mechanisms in debris flows at mt. Thomas, New Zealand, and implications for flow mobility. Sed., 28, 49-60.

PIERSON, T.C. & SCOTT, K.M. (1985). Downstream dilution of a lahar: Transition from debris flow to hyperconcentrated streamflow. Water Resources Res., 21, 1511-1524.

PIPER, J.D.A. (1978). Palaeomagnetic survey of the (Palaezoic) Shelve inlier & Berwyn Hills, Welsh Borderlands. Geophys. J. Royal Astr. Soc., 53, 355-371.

PLASS, L. van der & POBI, A.C. (1965). A chart for judging the reliability of point counting results. Am. J. Sci., 263, 87-90.

PLINT, A.G. (1983). Sandy fluvial point bar sediment from the middle Eocene of Dorset, England. IAS 6, 355-368.

PLINT, A.G. & WALKER, R.G. (1987). Morphology of an erosion surface out into the bad heart formation during major sea level changes, Santonian of west-central, Alberta, Canada. J. Sediment. Petrol., 57, 639-650.

PLINT, A.G., WALKER, R.G. & BERGMAN, K.M. (1986). Cardium Formation 6. Stratigraphic framework of the Cardium in subsurface. Bulletin of Canadian Petroleum Geology, v. 34, p.213-225.

POLLARD, J.E., STEEL, R.J. & UNDERSRUD, E. (1982) Facies sequence and trace fossils in lacustrine fan delta deposits, Hornelen basin (Middle Devonian), Western Norway. Sedimentary Geology, 32, 63-87.

POSTMA, G. (1983) Water escape structures in the context of a depositional model of a mass flow dominated conglomeratic fan delta (Abrioja Formation, Pliocene, Almeria basin, SE Spain). Sedimentology, 30, 91-103.

POSTMA, G. (1984). Mass flow conglomerates in a submarine canyon: Abrioja fan delta, Pliocene, south-east Spain. In: Koster, E.H. & Steel, R.J. 1984. Sed. of gravels & conglomerates. Can. Soc. Petrol. Geo. Mem.10 p.237-258.

PRIOR, D.B., BORNHOLD, B.D. & JOHNS, M.W. (1986) Active sand transport along a fjord bottom channel, Bute Inlet, B.C.. Geology, 14, 581-584.

PUDSEY, C.J. (1984) Fluvial to marine transition in the Ordovician of Ireland - a humid-region fan-delta? Geological Journal., 19, 143-172.

PUIGDEFABREGAS, C. & VLIET, A. VON (1978). Meandering deposits from the tertiary of the southern Pyrenees. In: Fluvial sedimentology. (ed. by A.D. Miall) Can. Soc. Petrol. Geol Mem 5., p.469-486.

RAAF, J.F.M. de, BOERSMA, J.R. & VAN GELDER, A.V. (1977) Wave generated structures and sequences from a shallow marine succession, Lower Carboniferous, County Cork, Ireland. Sedimentology., 24, 451-483.

RAMOS, A. & SOPENA,A. (1983). Gravel bars in low sinuosity streams Permian Triassic, central Spain. In: Modern and Ancient fluvial systems (ed. by J.D. Collinson and J. Lewin). Spec. Pub. Int. Assoc. Sediment., 6,, 301-312.

RAMPINO, M.R. & SANDERS, J.E. (1980) Holocene transgression in south-central Long Island, New York. J. Sediment. Petrol., 50, 1063-1079.

RAMSAY, A.C. (1866) The geology of North Wales. Mem. Geol. Surv. G.B. III.

RAMSAY, A.C.(1881) 2nd Ed. The geology of North Wales, Mem. Geol. Surv. G.B., vol. 3

RAST, N. (1961) Mid-Ordovician structures in SW Snowdonia. Liverpool and Manchester Geological Journal., 2, 645-652.

RAST, N. (1961) Mid-Ordovician structures in south-western Snowdonia. Lpool and Manchr. geol. J., 2, 645-652.

RAST, N. (1969) The relationship between Ordovician Structure and volcanicity in Wales. In: "The Precambrian and Lower Palaeozoic Rocks of Wales" (ed. by A. Wood), Univ. of Wales Press, p.305-335.

RAST, N., BEAVON, R.V., & FITCH, F.J. (1958) Subaerial volcanicity in Snowdonia. Nature. London., 181, 508-509.

RAST, N., BEAVON, R.V., and F.J. FITCH (1958) Sub-aerial volcanicity in Snowdonia. Nature, 181, p. 508.

RAST, N., FITCH, F.J. & BEAVON, R.V. (1954) The nomenclature and diagnostic characters of ignimbrites, with reference to Snowdonia. Abstr. geol. Soc. London, no. 1571, 116-121.

READING, H.G. (1986) African Rift tectonics and sedimentation, an introduction. In: Sedimentation in the African Rifts. (ed. by L.E. Frostick,), Geological Society Special Publication No. 25, 3-7.

READING, H.G. (1986) Problems and perspectives. In: Sedimentary Environments and Facies (Ed. by H.G. Reading), p. 520-524, Blackwell Scientific Publications, 615 pp.

REEDMAN, A.J., COLMAN, T.B., CAMPBELL, S.D.G. & HOWELLS, M.F. (1985) Volcanogenic mineralization related to the Snowdon volcanic Group. J. Geol. Soc. London., 142, 875-888.

REEDMAN, A.J., HOWELLS, M.F., ORTON, G.J. & CAMPBELL, S.D.G. (1987). The pitts head tuff formation : a subarial to submarine welded ash flow tuff of Ordovician age, north Wales. Geol.Mag., 124, 427-439.

REEDMAN, A.J., LEVERIDGE, B.E., & EVANS, R.B. (1984) The Arfon Group ('Arvonian') of North Wales. Proc. Geol. Assoc., 95, 313-321.

REID, I & FROSTICK, L.E. (1985) Beach orientation, bar morphology and the concentration of metalliferous placer deposits: a case study, Lake Turkana, N. Kenya. J. Geol. Soc. London., 142, 837-848.

REILINGER, R. & OLIVER, J. (1976) Modern uplift associated with a proposed magma body in the vicinity of Soccorro, New Mexico., Geology, 4, 583-586.

REILINGER, R.E. & YORK, J.E. (1979). Relative crustal subsidence levelling data in a seismically active part of the Rio Grande Rift, New Mexico. Geology., 7, 139-143.

REILINGER, R.E., CITRON, G.P. & BROWN, L.D. (1977). Recent vertical crustal movements from precise levelling data in southwestern Montana, western Yellowstone National Park, and the Snake River plain. J.G.R., 82, 5349=5359.

REINSON, G.E. (1984). Barrier Island and associated strand-plain systems. In: Walker, R.G. (ed.) Facies Models. Geoscience Canada. Reprint series 1. p.119-140.

RHODES, E.G. (1982) Depositional model for a chenier plain, Gulf of Carpentaria, Australia. Sedimentology, 29, 201-221.

RINE, J.M., HELMHOLD, K.P., & BARTLETT, G.A. (1987) Discussion of: Cardium Formation 6: Stratigraphic framework for the Cardium in subsurface. Bull. Can. Petrol. Geol., 35, 362-363.

ROBBINS, E.I. (1983) Accumulation of fossil fuels and metallic minerals in active and ancient rift systems. Tectonophysics, 94, 633-658.

ROBERTS, B. (1967) Succession and structure in the Llwyd Mawr syncline, Caernarvonshire, North Wales. Geol. J., 5, 369-390.

ROBERTS, B. (1969) The Llwd Mawr Ignimbrite and its associated volcanic rocks. In: "The Precambrian and Lower Palaeozoic Rocks of Wales" (ed. by A. Wood), Univ. of Wales Press, p. 337-356.

ROBINSON, D. & BEVINS, R.E. (1985) Incipient metamorphism in the Lower Palaeozoic marginal basin of Wales. J. Metamorphic Geol., 4, 101-114.

ROEP TH. B., BEETS, D.J., DRONKERT, H. and PAGNIER, H. (1979). A prograding coastal sequences of wave built structures of Messinian age, Sorbas, Almeria, Spain, Sedimen. Geol., 22, 135-163.

ROGERS, P. & GUEST, J.E. (1967) The sintering of glass and its relationship to welding in ignimbrites. Proc. geol. soc. London., 1641, 174-177.

SHAW, H.R. (1974) Obsidian-H_2O viscosities at 1000 and 2000 bars in the temperature range 700 to 900°C. J. Geophys. Res., 68, 63337-6343.

RONOV, A.B. (1964). Common tendencies in the chemical evolution of the earths crust, ocean and atmosphere. Geokhimiya, 8, p.715-743.

ROOBOL, M.J. (1976) Post-eruptive mechanical sorting of pyroclastic material - an example from Jamaica. Geol. Mag., 113, 429-440.

ROSS, G.M. & SMITH, T.L. (1985) Physical volcanology and stratigraphy of the Seton Formation: foredeep volcanism in the Athapuscow Aulacogen, District of Mackenzie. Current research, Part A. GSC Paper 85-1A, 681-692.

ROYDEN, L. & KEEN, C.E. (1980) Rifting process and thermal evolution of the continental margin of eastern Canada, determined from subsidence data. Earth. Planet. Sci. Letters, 5, 343-361.

RUBIN, D.M. (1987) Formation of Scalloped cross-bedding without unsteady flows. J. Sediment. Petrol., 39-45.

RUST, B.R. (1972) Structures and process in a braided river. Sedimentology, 18, 221-245.

RUST, B.R. (1978). Depositional models for braided alluvium. In: "Fluvial sedimentology", (Ed.. by A.D. Miall). Can. Soc. Petrol. Geol. Mem. 5 p.605-627.

RUST, B.R. (1979) Coarse alluvial deposits. In: Facies Models, (ed. by R.G. Walker), Geoscience Canada Reprint series 1, second edition, 53-70.

RUST, B.R. & KOSTER, E.H. (1984) Coarse alluvial deposits. In: Facies Models, (ed. by R.G. Walker), Geoscience Canada Reprint Series 1, second edition, 53-70.

RUTLEY, F. (1881). The microscopical structure of Devitrified rocks from Beddgelert & Snowdon. Quart. J. Geol. Soc. London, v. xxxvii, 403-409.

RYDER, J.M. (1971) Some aspects of the morphometry of paraglacial alluvial fans in south central British Columbia. Can. J. Earth Sci., ?, 1252-1264.

RYTUBA, J.J. & McKEE, E.H. (1981). Peralkaline ash flow tuffs and calderas of the McDermitt volcanic field southeast Oregon and north central Nevada. J.G.R., 89 Bio, 8616-8627.

SARNA-WOJCICKI, A.M., MORRISON, S.D., MEYER, C.E. & HILLHOUSE, J.W. (1987). Correlation of upper cenozoic lephra layers between sediments of the western United States and eastern Pacific ocean and comparison

with biostratigraphic and magnetostratigraphic ice age data. Bull. Geol. Soc. Am., 98, 207-223.

SAUNDERS, A.D. & TARNEY, J. (1984) Geochemical characteristics of basaltic volcanism within back-arc basins. In: "Marginal Basin geology: volcanic and associated sedimentary and tectonic processes in modern and ancient marginal basins. (ed. by B.P. Kokelaar and M.F. Howells), p. 59-76.

SAVAGE, J.C. & CLARK, M.M. (1982). Magmatic resurgence in the Long Valley caldera, California - possible cause of the 1980 Mammoth Lakes. EQ. Science., 217, 531.

SCHERMERHORN, L.J.G. (1970) The deposition of volcanics and pyrite in the Iberian Pyrite belt. Mineral. Deposita, 5, 273-279.

SCHERP, A. & GRABERT, H. (1983) Unterdevonische schmelzetuffe im rechtsrheinischen schiefergebirge. N. Jb. Geol. Pal. Mh., xx, 47-58.

SCHMID, R. (1981). Descriptive nomenclature and classification of pyroclastic deposits ans fragments: Recommendations of the IUGS Subcommision on the Systematics of Igneous Rocks. Geology 9, 41-43.

SCHRAMM, W.E. (1981) Humid tropical alluvial fans, northwest Honduras. Unpub. M.Sc. thesis. Louisiana State University.

SCHULTZ, A.W. (1982) The deposits, internal structure and geometry in six alluvial fan-fan delta bodies (Devonian), Norway - a study in the significance of bedding sequence in conglomerates - a discussion. J. Sediment. Petrol., 53, 325-347.

SCHUMM, S.A. (1963). Sinuosity of alluvial rivers in the Great Plains. G.S.A. Bull. 74, 1089-1100.

SCHUMM, S.A. (1981). Evolution and response of the fluvial system sedimentological implications. In: Ethridge, F.G. and Flores, R.M. Recent and Ancient Nonmarine Depositional Enviroments: Models for exploration. SEPM., 31, 19-29.

SCHWAB, W.C. & Lee, H.J. (1988). Causes of two slope failure types in continental shelf sediment, northeastern gulf of Alaska. J. Sediment. Petrol., 58, 1-11.

SCLATER, J.G., ROYDEN, L., HORVATH, F. BURCHFIEL, B.C., SEMKEN, S. & STEGENA, L. (1980) The formation of the intra-Carpathian basins as determined from subsidence data. Earth Planet. Sci. Letters, 51, 139-162.

SEDGEWICK, A. (1838) Synopsis of the English series of the stratified rocks Inferior to the Old Red Sandstone. Proc. Geol. Soc., ii, 679-xxx.

SEDGEWICK, A. (1847) On the classification of fossiliferous slates of North Wales, Cumberland, Westmoreland and Lancashire. Q. J. Geol. Soc. London, xx. 133-164.

SELF, S. & RAMPINO, M.R. (1981). The 1883 eruption of Krakatau. Nature 294, 699-704.

SELF, S. & SPARKS, R.S.J. (1978). Characteristics of wide-spread

pyroclastic deposits formed by the interaction of silicic magma and water . Bull. Volcanol. 41-3, 1-17. 4,6,7,9

SELF, S., RAMPINO, M.R., NEWTON, M.S. & WOLFF, J.A. (1984). Volcanological study of the great Tambora eruption of 1815. Geology v 12 p.659-663.

SEYMOUR, R.J. (1986). Nearshore auto suspending turbidity flows. Ocean Engineering., 13, 435-447.

SHACKLETON, R.M. (1954) The structural evolution of North Wales. Liverpool Manchr. geol. J., 2, 216-252.

SHACKLETON, R.M. (1959) The stratigraphy of the Moel Hebog district between Snowdon and Tremadoc. Liverppol and Manchester. Geol. Journal, 2, 216-252.

SHAW, H.R. (1980). The fracture mechanisms of magma transport from the mantle to the surface. In: Physics of Magmatic processes (ed.by. R.B. Hargrayes) Princeton Univ.Press. Princeton, NJ.

SHENHAV, H. (1971). Lower cretaceous sandstone reservoirs, Israel, petrogrophy, porosity, permeability. Am. Assoc. Petrol. Geol., 55, 2194-2224.

SHORT, A.D. & HESP, P.A. (1981) Wave, beach and dune interactions in southeastern Australia. Marine Geology, 48, 259-284.

SIGURDSSON, H., CAREY, S.N. & FISHER, R.V. (1987). The 1982 eruptions of El Chichon volcano, Mexico (3): Physical properties of pyroclastic surges. Bull. Volcanol., v49, p.467-488.

SIGURDSSON, H., SPARKS, R.S.J., CAREY, S.N. & HUANG, T.C. (1980). Volcanogenic sedimentation in the Lesser Antilles Arc. J. Geol. 88, 523-540.

SIMPSON, J.E. (1972). Effect of lower boundary on the head of a gravity current. J. Fluid Mechanics, 53, 759-768.

SIMPSON, J.E. & BRITTER, R.E. (1979). The dynamics of the head of a gravity current over a horizontal surface. J. Fluid Mechanics, 94, 477-495.

SMITH, D.G. (1983). Anastamosed fluvial deposits: modern examples from western Canada. In: Modern and Ancient fluvial systems (ed. by J.D. Collinson and J. Lewin). Spec. Pub. Int. Assoc. Sediment., 6,, 155-168.

SMITH, G.A. (1986) Coarse grained non-marine volcaniclastic sediment: terminology and depositional process. Bull. Geol. Soc. Am., 97, 1-10.

SMITH, G.A. (1987) Volcanism-induced aggradation in fluvial basins: examples from the Pacific Northwest USA. In: Developments in fluvial sedimentology. (ed. by R. Flores, F. Ethridge, and M. Harvey). SEPM Sp. Pub. xx

SMITH, M. (1988) Tectonic evolution of Cambrian & Ordovician rocks in South Central Snowdonia. Unpublished doctoral thesis, Univ. Wales, Aberystwyth.

SMITH, N.D. (1971). Pseudo planar stratification produced by very low amplitude sand waves. J. Sediment. Petrol., 41, 69-73.

SMITH, N.D. (1972) Some sedimentary aspects of planar cross-stratification in a sandy braided river. J. Sediment. Petrol., 42, 624-634.

SMITH, N.D. (1972). Some sed aspects of planar cross stratification in a sandy braided river. J. Sediment. Petrol., 42, 624-634.

SMITH, R.L. (1960). Ash flows. Geol.Soc.Amer.Bull., 71, 795-842.

SMITH, R.L. & BAILEY, R.A. (1968). Resurgent cauldrons. Bull. Geol. Soc. Am., 6, 613-662.

SNEH, A. (1979) Late Pleistocene fan deltas along the Dead Sea rift. J. Sediment. Petrol., 49, 541-552.

SOLOMAN, M. & WALSHE, J.L. (1981) An investigation into the formation of volcanogenic hosted Mt. Lyell Copper deposits using geology, minerals, stable isotopes, and a six component chlorite solid solution model. Economic Geology, 76, 246-284.

SOLOMAN, M. & WALSHE, J.L.(1979) The Formation of massive sulphide deosits on the sea floor. Econ. Geol., 74, 797-813.

SOPER, N.J. and HUTTON, D.H.W. (1984) Late Caledonian sinistral displacements in Britain: implications for a three plate collision model. Tectonics, 3, 781-794.

SOPER, N.J., WEBB, B.C. and WOODCOCK, N.H. (1987) Late Caledonian (Acadian) transpression in north-west England: timing, geometry and geotectonic significance. Proc. York. Geol. Soc., 46, 175-192

SPARKS, R.S.J. (1976). Grain size variations in ignimbrites and implications for the transport of pyroclastic flows. Sedimentology., 23, 147-188.

SPARKS, R.S.J. & WALKER, G.P.L. (1973). The ground surge deposit: a third type of pyroclastic rock. Nature 241, 62-64.

SPARKS, R.S.J. & WILSON, L. (1976). A model for the formation of ignimbrite by gravitational column collapse. J. Geol. Soc. London 132, 441-451.

SPARKS, R.S.J., SIGURDSSON, H. & CAREY, S.N. (1980) The entrance of pyroclastic flows into the sea. I. Oceanographic and geological evidence from Dominica, Lesser Antilles. J. Volcanol. Geotherm. Res., 87-96.

SPARKS, R.S.J., SIGURDSSON, H. & CAREY, S.N. (1980) The entrance of pyroclastic flows into the sea.2. Theoretical considerations on subaqueous emplacement and welding. J. Volcanol. Geotherm. Res., 7, 97-105.

SPJELDNAES, N. (1985) Biostratigraphy of the Scandanavian Caledonides. In: The Caledonide orogen- Scandanavia and related areas. (ed. by D.G. Gee and B.A. Sturt) John Wiley and sons, Chichester. p. 317-330.

STANTON, W.I. (1960) The Lower Palaeozoic rocks of south west Murrisk, Ireland. Q. J. Geol. Soc. London, 116, 269-296.

STEEL, R.J. (1976) Devonian basins of western Norway: sedimentary response to tectonism and varying tectonic context. Tectonophysics, 36, 207-224.

STEEL, R.J. & AASHEIM, S.M. (1978) Alluvial sand deposition in a rapidly subsiding basin, (Devonian, Norway). In: Fluvial Sedimentology. (ed. by A.D. Miall). Can. Soc. Petrol. Geol. Mem. 5, 386-412.

STEPHENS, M.B. & GEE, D.G. (1985) A tectonic model for the evolution of the eugeoclinal terranes in the central Scandanavian Caledonides. In: The Caledonide orogen- Scandanavia and related areas. (ed. by D.G. Gee and B.A. Sturt) John Wiley and sons, Chichester. p. 953-978.

STILLMAN, C.J. (1984) Ordovician volcanicity. In: "Aspects of the Ordovician System" (ed. by D.L. Bruton). Palaeontological Contributions from the University of Oslo no. 295, Universitetsforlaget, p. 183-194.

STOLZ, J.F., CHANG, S-B.R., & KIRSCHVINK, J.L. (1986) Bacterial magnetite as trace fossil and paleooxygen indicator. In: Abstracts, 12th International Association of Sedimentologists Congress, 1986, Canberra, Australia. , 288.

STUBBLEFIELD, W.L., McGrail, D.W., & Kersey, D.G. (1984). Recognition of transgressive and post transgressive sand ridges on the New Jersey continental shelf. In: Siliclastic shelf sediments. (ed. by R.W. Tillman & C.T. Siemers). SEPM SP34 p.1-25.

STURM, M. & MATTER, A. (1978). Turbidites and varves in Lake Brienz, Switzerland: deposition of clastic detritus by density currents. In: Modern & Ancient Lake Sediments, (ed. by A. Matter & M. Tucker). IAS SP2. p.145-166.

SUCZEK, C.A. & INGERSOLL, R.V. (1985) Petrology and provenance of Cenozoic sand from the Indus cone and the Arabian basin, DSDP sites 221, 222, 224. J. Sediment. Petrol., 55, 340-346.

SURDAM, R.C. & BOLES, J.R. (1979) Diagenesis of volcanic sandstones. In: "Aspects of diagenesis" (ed. by P.A. Scholle and P.R. Schluger) SEPM Spec. Pub. 26, p.227-242.

SURLYK, F. (1984) Fan-delta to submarine fan conglomerates of the Volgian-Valanginian Wollaston foreland group, East Greenland. In: Sedimentology of gravels and conglomerates. (ed. by E.H. Koster and R.J. Steel) Can. Soc. Petrol. Geol. Mem. 10, 359-382.

SURLYK, F. & NOE-NYGAARD, N. (1986). Hummocky cross-stratification from the Lower Jurassic Hasle Formation of Bornholm, Denmark: Sedimentary Geology, v.46, p.259-273.

SUTTNER, L.F. & BASU, A. (1985) The effect of grain size on detrital modes: a test of the Gazzi-Dickinson point counting method. Discussion. J. Sediment. Petrol., 55, 616-617.

SUTTNER, L.J. & DUTTA, P.K. (1986) Alluvial sandstone composition and paleoclimate: I: Framework mineralogy. J. Sediment. Petrol., 56, 329-345.

SWANBERG, C.A. (1983) Geothermal resources of rifts: a comparison of the

Rio Grande and Salton trough. Tectonophysics, 94, 659-678.

SWANSON, E.R. & McDOWELL, F.W. (1985) Geology and geochronology of the Tomochic caldera, Chihuahau, Mexico. Geol. Soc. Am., 96, 1477-1482.

SWIFT, D.J.P. & NUMMEDAL, D. (1987) Hummocky cross stratification, tropical hurricanes and intense winter storms: discussion. Sedimentology., 34, 338-344.

SWIFT, D.J.P., FIGUEIREDO, A.G., FREELAND, G.L. & OERTEL, G.F. (1983). Hummocky cross sratification and megaripples: a geological double standard. J. Sediment. Petrol., 53, 1295-1317.

SYKES, R.M. & BRAND, R.P. (1976) Fan delta sedimentation: an example from the late Jurassic-early Cretaceous of Milneland, Central East Greenland. Geologie en Mijnbouw, 55, 195-203.

SYVITSKI, J.P.M. & FARROW, G.F. (1983) Structures and processes in bayhead deltas: Knight and Bute Inlet, British Columbia. Sedimentary Geology., 36, 217-244.

TANKARD, A.J. & BARWIS, J.H. (1982). Wave dominated deltaic sedimentation Devonian Bokkerel basin of South Africa. J. Sediment. Petrol., 52 959-974.

THACKRAY, J.C. (1976) The Murchison-Sedgewick controversy. J. Geol. Soc. London, 132, 367-372.

THOMAS, C. & BRIDEN, J.C. (1976) Anomalous geomagnetic field during the late Ordovician. Nature, 259, 380-382.

THORARINSSON, S. (1954). The eruption of Hekla 1947-48, part 2, Ch.3, The tephra-fall from Hekla on March 29, 1947. Soc. Sci. Islandica, Reykjavik, 1-68.

THORPE, R.S. (1974) Aspects of magmatism and plate tectonics in the Precambrian of England and Wales. Geol. Journal, 9, 115-136.

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. London, 141, 521-536.

THORPE,R.S. (1974) Aspects of magmatism and plate tectonics in the Precambrian of England and Wales. Geological Journal, 9, 115-136.

THURSTON, P.C. (1980) Subaerial volcanism in the Archean Uchi-Confederation volcanic belt. Precambrian Research, 12, 79-80.

TILLMAN, R.W. (1985). A spectrum of shelf sands and sandstones. In: Shelf sands & sandstone reservoirs (ed. by R.W. Tillman, D.J.P. Swift, & R.G. Walker). SEPM short course notes no.13, p.1-46.

TILLMAN, R.W. & Martinsen, R.S. (1984). The Shannon shelf ridge sandstone complex, salt over anticline area, powder river Basun, Wyoming. In: Siliclastic shelf sediments. (ed. by R.W. Tillman & C.T. Siemers). SEPM SP34 p.85-142.

TRYTHALL, R.J.B., ECCLES, C., MOLYNEUX, S.G. & TAYLOR, W.E.G. (1987). Age and controls of ironstone deposition (Ordovician) North Wales. Geol.

J., 22, 31-44.

TUNBRIDGE, I.P. (1981). Sandy high energy flood sedimentation-some criteria for recognition with an example from the Devonian of South-west England. Sediment.Geol., 28, 79-95.

TYLER, J.E. & WOODCOCK, N.H. (1987). The Bailey Hill Formation: Ludlow Series turbidites in the Welsh Borderland reinterpreted as distal storm deposits. Geol. J., 22, 73-86.

VALENTINE, G.A. (1987). Stratified flow in pyroclastic surges. Bull. Volcanol., v49, p.616-630.

VALENTINE, G.A. & FISHER, R.V. (1986). Origin of layer 1 deposits in ignimbrites., Geology., 14, 146-148.

VALLONI, R. & MAYNARD, J.B. (1981): Detrital modes of recent deep-sea sands and their relation to tectonic setting: a first approximation; Sedimentology, 28, 75-80.

VALLONI, R. & MEZZADRI, G. (1984) Compositional suites of terrigenous deep sea sands of the present continental margins. Sedimentology, 31, 353-364.

VAN der MEULEN, S. (1983) Internal structure and environmental reconstruction of Eocene transitional fan delta deposits, Monilobat-Castigaleu formations, Southern Pyrenees, Spain. Sedimentary Geology, 37, 85-112.

VAUGHAN-THOMAS, W. (1985). Wales: A History. Michael Joseph, London, 269 pp.

VELBEL, M.A. (1985) Mineralogically mature sandstones in accretionary prisms, J. Sediment. Petrol., 55, 685-690.

VESSEL, R.K. & DAVIES, D.K. (1981) Non-marine sedimentation in an active forearc basin. In: "Non-marine depositional environments" (ed. by F.G. Ethridge and R.M. Flores), SEPM Spec. Pub. 31, 31-45.

VOS, R.G. (1981) Sedimentology of an Ordovician fan delta complex, western Libya. Sedimentary Geology, 18, 161-173.

VOS, R.G. & ERIKSSON, K.A. (1977) An embayment model for tidal and wave swash deposits occurring within a fluvially dominated middle Proterozoic sequence in South Africa. Sedimentary Geology., 18, 161-173.

WALKER, G.P.L. (1972) Crystal concentration in ignimbrites. Contr. Mineral. Petrol., 36, 135-146.

WALKER, G.P.L., (1972). Crystal concentration in ignimbrites. Contr. Mineral. Petrol. 36, p. 135-146.

WALKER, G.P.L. (1979) A volcanic ash generated where ignimbrite entered the sea. Nature, 281, 642-646.

WALKER, G.P.L. (1985). Origin of coarse lithic breccias near ignimbrite source vents. J. Volcanol. Geotherm. Res., 25, 157-171.

WALKER, J. (1977). Drops of water dance on a hot skillet and the

experimenter walks on hot coals. Sci. Am. (2), 126-130.

WALKER, R.G. (1984) Shallow marine environments. In: Facies Models (ed. by R.G. Walker). Geoscience Canada Reprint Series.

WALKER, R.G. (1985). Geological evidence for storm transportation and deposition on ancient shelves. In: Shelf sands & sandstone reservoirs (ed. by R.W. Tillman, D.J.P. Swift, & R.G. Walker). SEPM short course notes no.13 p. 243-302.

WALKER, R.G. & HARMS, J.C. (1971) The Catskill delta: a prograding muddy shoreline in central Pennsylvania. J. Geology, 79, 381-399.

WALKER, R.G., DUKE, W.L. & LECKIE, D.A. (1983) Hummocky cross-stratification: significance of its variable bedding sequence: Discussion. Bull. Geol. Soc. Am., 94, 1245-1249.

WALKER, T.R. (1984) 1984 SEPM presidential address: diagenetic albitization of Potassium feldspar in arkosic sandstones. J. Sediment. Petrol., 54, 3-16.

WALKER, T.R., WAUGH, B. & CRON, A.J. (1978). Diagenisis of first cycle desert alluvium of cenozoic age, south western United States and north western Mexico. Bull. Geol. Soc. Am., 89, 19-32.

WALTON, A.W. (1977) Petrology of volcanic sedimentary rocks, Vieja Group, Southern Rim, Rock county, Trabs Pecos, Texas. J. Sediment. Petrol., 47, 137-157.

WALTON, A.W. (1979) Volcanic sediment apron in the Tascotal formation (oligocene?), Trans-pecos, Texas. J. Sediment. Petrol., 49, 303-314.

WALTON, A.W. (1986) Contrasting sedimentation in continental volcanic terranes: Trans-Pecos Texas (Eocene-Oligocene) and Marysvale, Utah (Oligocene-Miocene), USA. Abstract. 12 th. Assoc. Sedimentologists, Congress, 315.

WALTON. A.W. (1986) Effect of Oligocene volcanism on sedimentation in the Trans-Pecos volcanic field of Texas. Bull. Geol. Soc. Am., 97, 1192-1207.

WARD, J.C. (1875) Notes on the comparative microscopic rock structure of some ancient and modern volcanic rocks. Q. J. Geol. Soc., 31, 388-422.

WARD, L.G., STEPHEN, M.F. & NUMMEDAL, D. (1976) Hydraulics and morphology of glacial outwash distributaries, Skeidararsandur, Iceland. J. Sediment. Petrol., 46, 770-777.

WASSON, P.J. (1977) Catchment processes and the evolution of alluvial fans in the lower Derwent valley, Tasmania. Z. Geomorph., N.J., 21, 147-168.

WATSON, J. & DUNNING, F.W. (1979) Basement cover relations in the British Caledonides. In: "The Caledonides of the British Isles. Geol. Soc. London., Special Publication 16, p. 67-92.

WELLS, J.T. & COLEMAN, J.M. (1981) Physical processes and fine grained sediment dynamics, Coast of Surinam, South America. J. Sediment. Petrol., 51, 1053-1068.

WELLS, N.A. (1983) Transient streams in sand poor redbeds: early-middle Miocene Kuldana formation of northern Pakistan. In: Modern and Ancient fluvial systems (ed. by J.D. Collinson and J. Lewin). Spec. Pub. Int. Assoc. Sediment., 6,, 393-403.

WERNICKE, B.(1985) Uniform sense simple shear of the continental lithosphere. Can. J. Earth Sci., 22, 108-125.

WESCOTT, W.A. & ETHRIDGE, F.G. (1982) Bathymetry and sediment dispersal dynamics along the Yallahs fan delta front, Jamaica. Marine Geology, 46, 245-260.

WESCOTT, W.A. & ETHRIDGE, F.G. (1980) Fan delta sedimentology and tectonic setting - Yallahs fan delta, southeast Jamaica. Bull. Am. Assoc. Petrol. Geol., 64, 374-399.

WESTGATE, J.A. & GORTON, M.P. (1981). Correlation techniques in tephra studies. In: Self, S. & Sparks, R.S.J. (eds)., Tephra studies. D. Reidel Publ. Co., Dordecht. Holland, 73-94.

WHITE, R.S., SPENCE, G.D., FOWLER, S.R., McKENZIE, D.P., WESTBROOK,G.K. & BOWEN, A.N. (1987). Magmatism at lifted continental margins. Nature 330, 6147, 439-444.

WHITTINGTON, H.B., DEAN, W.T., FORTEY, R.A., RICKARDS, R.B., RUSHTON, A.W.A., & WRIGHT, A.D. (1984). Defination of the Tremadoc series and the series of the Ordovician system in Britain. Geol. Mag., 121, 17-33.

WILKINSON, I. (1987). A finite study of the Ordovician volcanic rocks of Snowdonia, North Wales, and its implications for a regional strain model. Geol. J., 22, 95-106.

WILKINSON, I. (1988) Deformation of the Ordovician volcanics of Snowdonian, North Wales. Unpublished doctoral thesis, Univ. Wales, Aberystwyth.

WILLIAMS, A. (1963) The Caradocian brachiopod faunas of the Bala district, Merionethshire. Bull. Br. Mus. Nat. Hist., 8, 327-472.

WILLIAMS, A. (1976). Plate tectonics and biofacies evolution as factors in Ordovician correlation. In: Basset, M.G. (ed). The Ordovician System: proceedings of a palaeontological association symposium, Birmingham, September 1974. Univ of Wales press. p.29-66.

WILLIAMS, D. (1930) The geology of the country between Nant Peris and Nant Ffrancon, Snowdonia. Quart. J. geol. Soc. London, 86, 191-230.

WILLIAMS, H. (1922) The igneous rocks of the Capel Curig District (North Wales). Proc. Liverpool Geol. Soc., 13, 166-202.

WILLIAMS, H. (1927) The geology of Snowdon. Quart, J. geol. soc. London, 83, 346-431.

WILLIAMS, H. & BULMAN, O.M.B. (1931) The geology of the Dolwyddelan syncline (North Wales). Quart. J. Geol. Soc. London, 87, 425-458.

WILLIAMS, L.A. & CRERAR, D.A. (1985) Silica diagenesis II General mechanisms. J. Sediment. Petrol., 55, 312-321.

WILLIAMS, L.A. PARKS, G.A. & CRERAR, D.A. (1985) Silica diagenesis I: Solubility controls. J. Sediment. Petrol., 55, 301-311.

WILSON, C.J.N. & WALKER, G.P.L. (1982) Ignimbrite depositional facies: the anatomy of a pyroclastic flow. J. geol. Soc. London, 139, 581-592.

WILSON, C.J.N. & WALKER, G.P.L. (1982). Ignimbrite depositional facies: the anatomy of a pyroclastic flow. J. Geol. Soc. 139, 581-592.

WILSON, G. (1982). Introduction to small scale geological structures. In: Allen & Unwin London, p.128.

WILSON, J.T. (1966) Did the Atlantic close and then reopen? Nature., 211, p.676.

WINCHESTER, J.A. & FLOYD, P.A. (1977). Geochemical discrimination of different magma series and their differentiation products using immobile elements. Chem Geol., 20, 325-343.

WINN, R.D. (1978). Upper Mesozoic plysch of Tierra del Fuego and south Georgia island: A sedimentological approach to lithospheric plate reconstruction. Bull. Geol. Soc. Am., 89, 533-547.

WINN, R.D. & DOTT, R.H. (J R) (1979). Deep water from channel conglomerates of later Cretaceous age, southern Chile. Sedimentology., 26, 203-228.

WOOD, A. & SMITH, A.J. (1959) The sedimentation and sedimentary history of the Aberystwyth Grits (Upper Llandoverian). Quart. J. Geol. Soc. London, 114, 163-195.

WOOD, J. (1980) Epiclastic sedimentation and stratigraphy in the North Spirit Lake and Rainy Lake areas: A comparison. Precambrian Research, 12, 227-255.

WOOD, M.L. & ETHRIDGE. F.G. (1988). Sedimentary architecture of Gilbertion and mouth bar type fan deltas, Paradox Basin, Colorado: implications for reservoir description.

WOODCOCK, N.H. (1984) Early Palaeozoic sedimentation and tectonics in Wales. Proc. Geol. Assoc., 95, 323-335.

WOODCOCK, N.H. (1984) The Pontesford Lineament, Welsh Borderland. J. Geol. Soc. London, 141, 1001-1014.

WRIGHT, D.K. (1979) A faunal analysis of Soudleyan/Lonvillian rocks in Snowdonia. Unpublished Phd. thesis, Kingston Polytechnic.

WRIGHT, J.V. & COWARD, M.P. (1977) Rootless vents in welded ash flow tuffs from Northern Snowdonia, Northern Wales, indicating deposition in a shallow water environment. Geol. Mag., 114, 133-140.

WRIGHT, L.D. (1977) Sediment transport and deposition at river mouths: a synthesis. Bull. Geol. Soc. Am., 88, 857-868.

WRIGHT, L.D. & COLEMAN, J.M. (1973) Variations in morphology of major river deltas as functions of ocean wave and river discharge regimes. Bull. Am. Assoc. Petrol. Geol., 57, 370-378.

WRIGHT, L.D. 1985. River deltas. In: Coastal sedimentary environments (ed. by R.A. Davis). Springer-Verlag, New York, p.1-76.

WRIGHT, L.D., CHAPPELL, I., THOM, B.G., BRADSHAW, M.P. & COWELL, M.P. (1979) Morphodynamics of reflective and dissipative beach and inshore systems, Southeastern Australia. Marine Geology, 32, 105-140.

WRIGHT, L.D., THOM, B.G. & HIGGINS, R.J. (1980) Wave influences on river mouth depositional processes: examples from Australia and Papua, New Guinea. Estuarine and Coastal Marine Science, 11, 263-277.

YAMADA, E. (1973). Subaqueous pumice flow deposits in the Onikobe Caldera, Miyagi Prefecture, Japan. J. Geol. Soc. Jap., 79, 585-597.

YAMADA, E. (1984) Subaqueous pyroclastic flows: their development and their deposits. In: "Marginal Basin Geology. Volcanic and associated sedimentary and tectonic processes in modern and ancient marginal basins." (ed. by B.P. Kokelaar and M.F. Howells) Geological Society Special Publication 16, 29-36.

YAMAMOTO, K. SUIGISAKI, R. & ARAI, F. (1986). Chemical aspects of the alteration of acidic tuffs and their application to siliceous deposits. Chemical Geology., 55, 61-76.

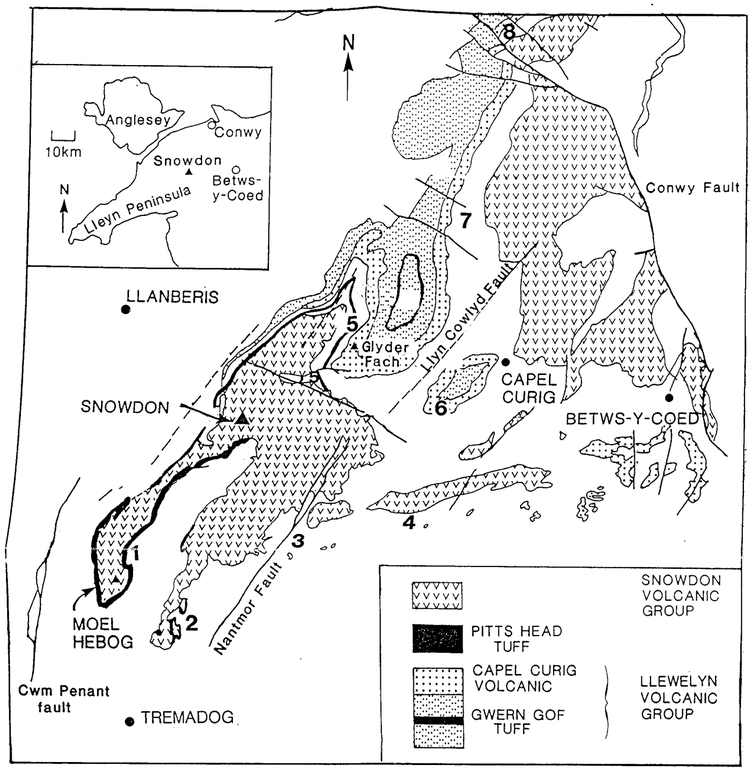
YERINO, L.N. & MAYNARD, J.B. (1984) Petrography of modern marine sands from the Peru-Chile trench and adjacent areas. Sedimentology, 31, 83-89.

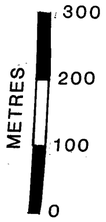
ZIEGLER, A.M. (1965) Silurian marine communities and their environmental significance. Nature, 207, 270-272.

ZIELINSKI, R.A. (1982). The mobility of uranium and other elements during alteration of rhyolite ash to montmorillonite: A case study in the Troublesome Formation, Colorado, U.S.A. Chemical Geology., 35, 185-204.

ENCLOSURE 1

STRATIGRAPHIC CROSS-SECTION ACROSS SNOWDONIA





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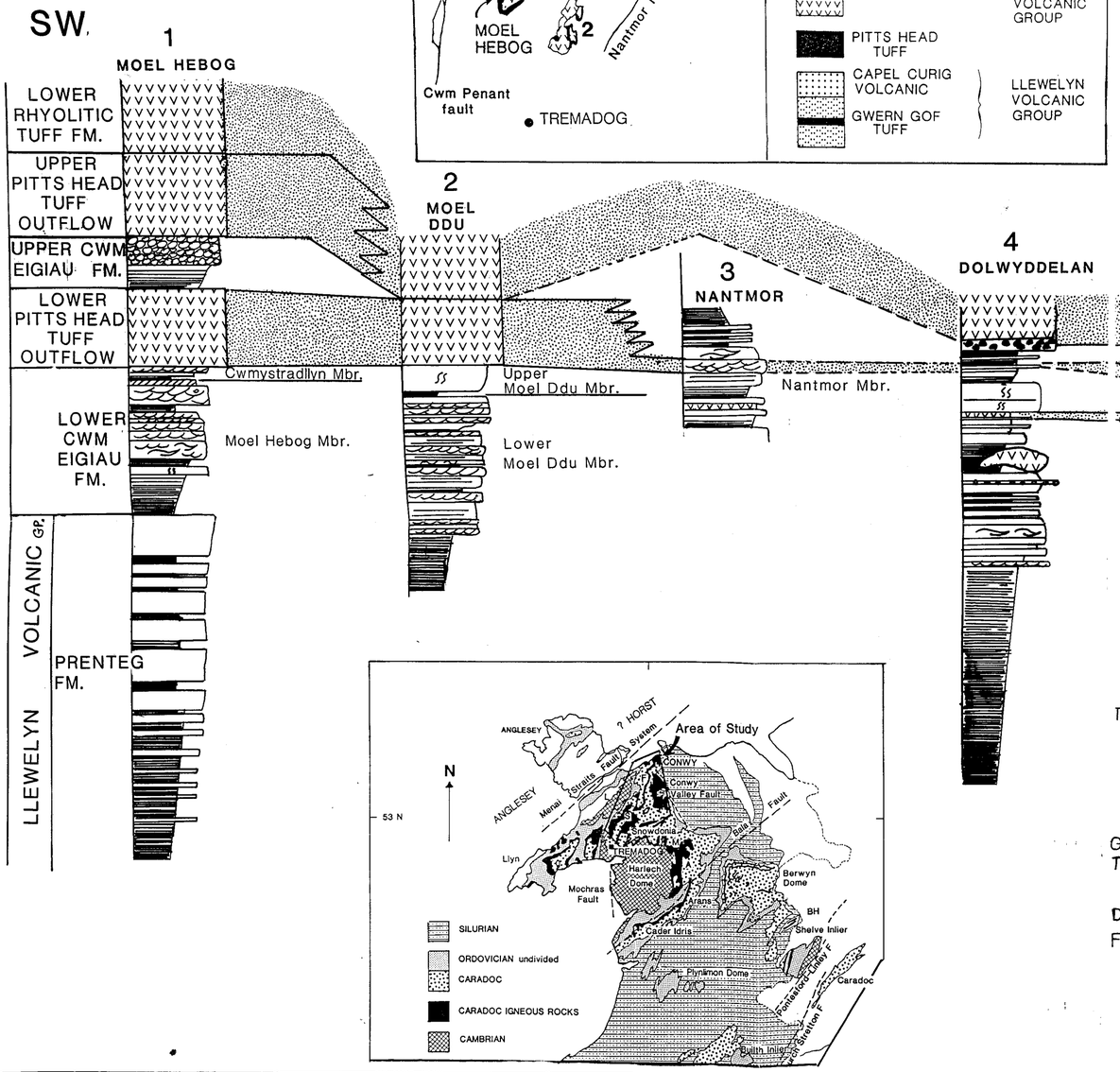

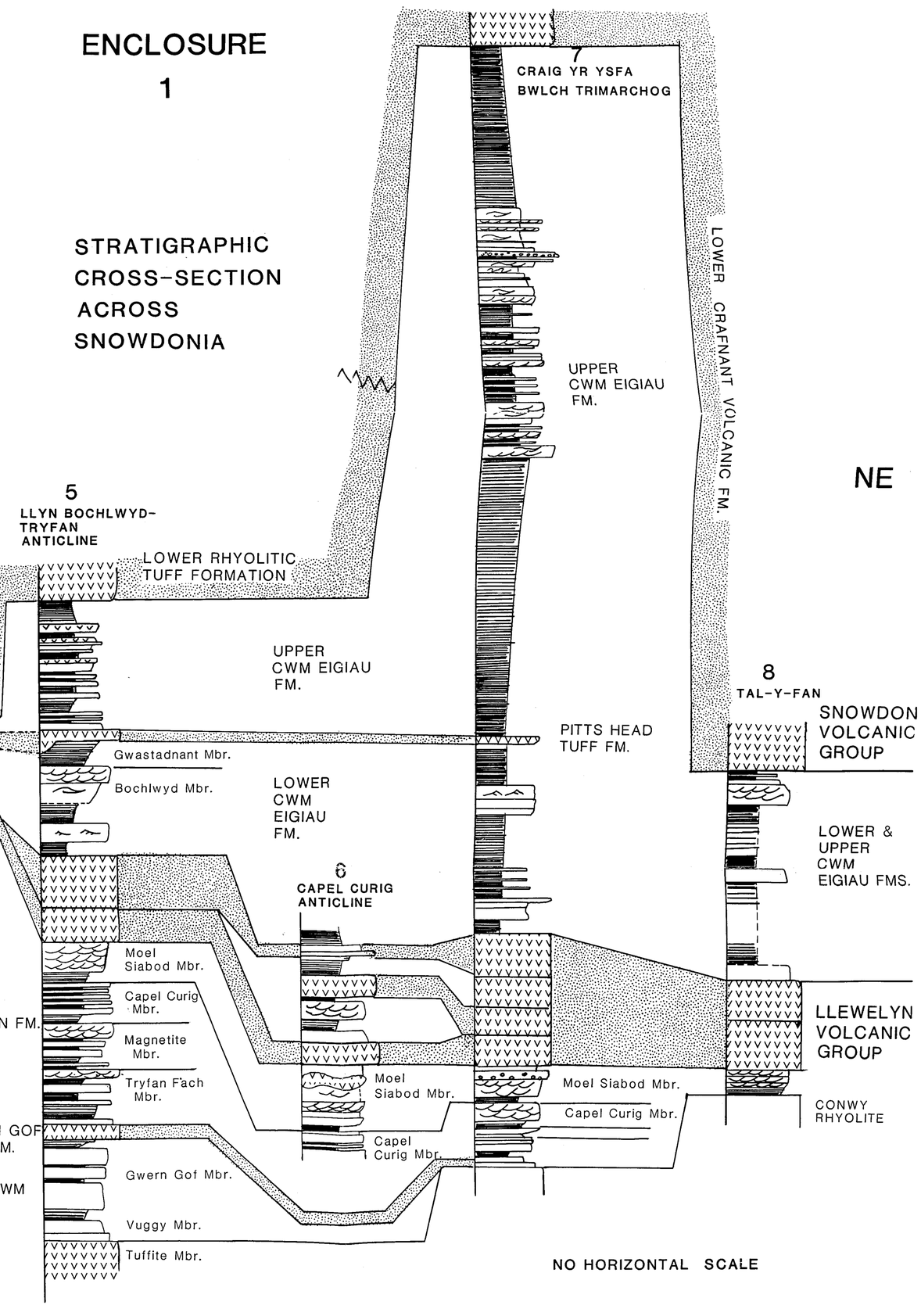

NE

NO HORIZONTAL SCALE

Journal of the Geological Society, London, Vol. **144**, 1987, pp. 523–525, 2 figs. Printed in Northern Ireland

ENCLOSURE 2

Discussion on the submarine eruption and emplacement of the Lower Rhyolitic Tuff Formation (Ordovician), North Wales *Journal*, Vol. 143, 1986, pp. 411–424

G. Orton writes: In a recent paper outlining the eruption and emplacement of the mid-Ordovician Lower Rhyolitic Tuff Formation (LRTF), Howells *et al.* (1986) invoke water depths of up to 500 m for the initial eruptions and a shallow marine environment for the second (main) phase of caldera development. Whilst not doubting that the second phase occurred in a shallow marine environment, I question their evidence for the initial phase submarine eruptions in considerable depths of water. The water depths associated with these initial eruptions have important implications regarding the formation of subaqueous *welded* ash flow tuffs.

The principal arguments of Howells *et al.* are the presence of marine sedimentary rocks and subaqueously emplaced basaltic lavas immediately subjacent to the LRTF, as well as the absence of significant Plinian fallout tuffs beneath ash flow deposits which they suggested may be a consequence of the suppression of the eruption column in substantial depths of water.

With regard to the depositional environments of the subjacent strata, the LRTF largely rests directly on the Pitts Head Tuff (PHT) to the SW of Snowdon. The absence of any intercalated sediments from Moel Hebog to Cwm Llan (Fig. 1), a distance of 10 km, suggests that this area was subaerial. The lack of significant erosion channels in the top of the PHT, such as one observes above subaerial ash flow tuffs on Tenerife would be a function of both the pervasive welding of the PHT and the absence of any significant intra-basinal topography; most pyroclastic volcanic fields are characterized by a gentle constructional topography (Lipman 1984). Pillow basalts/hyaloclastites associated with faulting at Cwm Llan are more likely to be related to the second stage of caldera development than to the initial eruptions.

Where a basin did exist to the NE, over 200 m of sediment was deposited between the PHT and the LRTF. At Pont y Gromlech (near the caldera margin), a fining upward sandstone sequence underlying the LRTF is interpreted as representing a transgressive, non-barred, wave influenced shoreline (Fig. 2). Fine sandstones immediately subjacent to the LRTF are dominantly parallel laminated to bioturbated, with minor swaley cross-stratification and wave rippling. This facies assemblage is typical of the lower shoreface-to-offshore transition zone (water depths of 9–18 m off California, Howard & Reineck 1981) or inner shelf (Bourgeois 1980). At Llyn Idwal, further to the NE, hummocky cross-stratified sandstones and tuffaceous turbidites interbedded with siltstone indicate deeper water. Water depth at the time of tuff eruption would be shallower than that indicated by the underlying sandstone facies as late stage (i.e. a few hundred years) pre-eruption uplift (e.g. at Mt St Helens; Moore & Albee 1981) would rarely be reflected in the sedimentary record. Any sedimentary evidence of late stage uplift would also be destroyed by tuff emplacement which commonly results in the fluidization (Kokelaar 1984) or liquefaction of 3–5 m of the underlying sediments (G. J. Orton personal observation).

The depths indicated by sedimentary facies are supported by brachiopod palaeocommunities. All shelly faunas at Pont-y-Gromlech indicate a Dinorthis–Macrocoelia community (S. Tunnicliffe, BGS internal report PD80/23). Although debris flows from volcanic islands can mix and transport depth controlled communities (Lockley 1984), the uniformity of faunal assemblages suggests little transport in this case. Brenchley & Cocks (1982) concluded that the depth spectrum of latest Ordovician communities found in the Oslo region, Norway was less than 100 m; the depths assigned by Pickerill & Brenchley (1976) of less than 25 m for the Caradocian Dinorthis–Macrocoelia community are not therefore unreasonable (W. S. McKerrow pers. comm. 1986).

In summary, the absence of sediments suggests that a flat or gently sloping 'plateau' or horst block existed in the SW of Snowdon before the first stage of caldera development. The Moel Hebog– Cwm Llan basal welded tuff sheet of the LRTF was thus deposited in a dominantly subaerial environment. This plateau was maintained after the PHT

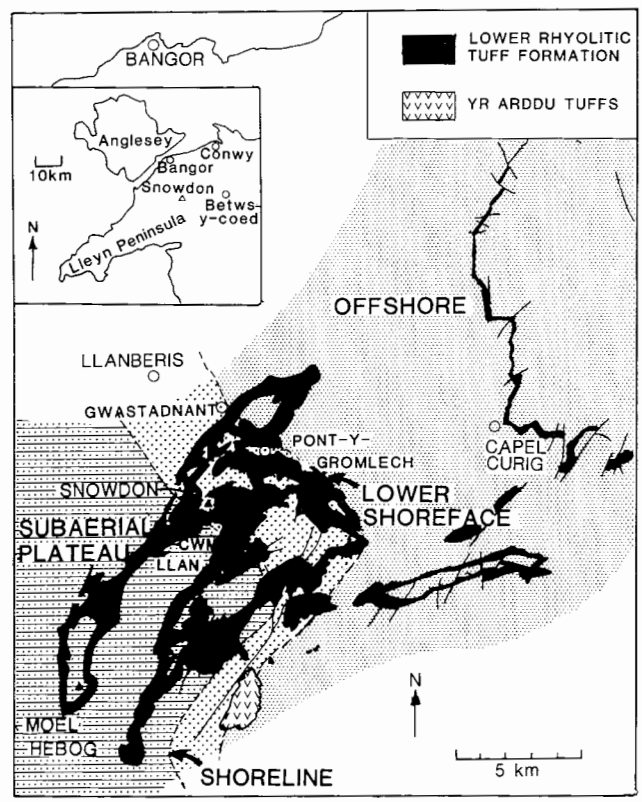

Fig. 1. Palaeogeography before initial eruptions of Lower Rhyolitic Tuff Formation.

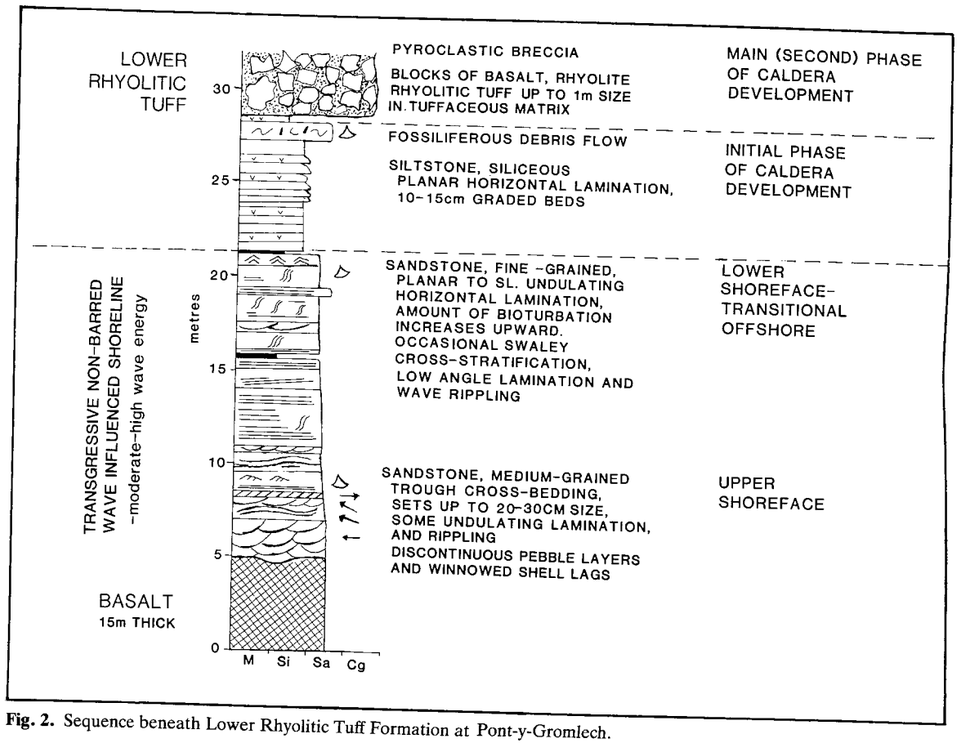

Fig. 2. Sequence beneath Lower Rhyolitic Tuff Formation at Pont-y-Gromlech.

eruption by buoyant forces related to the upward and/or lateral movement and accumulation of the LRTF magma body. The pre-eruption shoreline was probably near Snowdon itself as sedimentary facies and brachiopod palaeocommunities at Llanberis Pass in subjacent fine sandstones indicate maximum water depths of 25 m. The basin deepened to the east and NE. The caldera foundered only during its second stage with the development of the asymmetric submarine downsag volcanotectonic trough (Howells *et al.* 1986). The occurrence, however, of trough cross-bedding within outflow tuffs of the LRTF itself at Gwastadnant within their 'deeper marine basin', and hummocky cross-stratification within intracaldera tuffs on Moel Hebog (W. J. Fritz pers. comm. 1986) suggests that much of this sequence evolved above storm wave base as well, in water depths of less than 100 m.

A literature review (G. J. Orton in preparation) of the depositional environments of subaqueous welded tuffs suggests they are mostly confined either to extremely shallow water (less than flow thickness, e.g. Devonian, Germany, Scherp & Grabert 1981) from flows entering the sea or to deep water (greater than 500 m, e.g. Ordovician, South Wales, Kokelaar *et al.* 1985) where hydrostatic pressure confines submarine eruption columns and the increased solubility of water in glass decreases its viscosity (Sparks *et al.* 1980). Welded tuffs would be most unlikely at intermediate depths (50–500 m) such as those associated with the second phase of the LRTF caldera. Here, eruption columns would mix considerably with water resulting in cool volcaniclastic debris flows or high density turbidity currents (cf. Fiske & Matsuda 1964). With respect to ash flows entering the sea from land, changes in slope, or deflections from topographic barriers would result in incorporation of significant amounts of water through flow transformations (Fisher 1984). Any welding will be patchy and confined to 'hot spots' which ingested less water (e.g. Palaeogene, Phillipines; Fernandez 1969). More extensive welding at intermediate depths would only occur in special cases (e.g. Yr Arddu tuffs; Howells *et al.* in press) where the tuff is laterally restricted and the eruption column essentially falls back onto the vent.

M. F. Howells, A. J. Reedman and S. D. G. Campbell reply: Mr Orton's discussion of our paper is most welcome. In particular the summary (Fig. 1) of the local palaeogeography prior to the development of the eruption and emplacement of the LRTF adds valuable detail to our broadly similar, but more generalized model.

However, it is apparent from Orton's first paragraph that much of his critique is based on a misunderstanding of our views concerning water depths during the emplacement of the LRTF. We did not 'invoke water depths of up to 500 m for the initial eruptions and a shallow marine environment for the second (main) phase of caldera development'. We proposed (p. 421) an initial stage when 'a relatively small volume ($<10\ km^3$) was erupted in the central and southern areas' where it was reworked in 'a shallow marine to subaerial environment'. We also considered that, subsequent to this initial stage, the locus of the second, main caldera phase shifted northwards and caldera collapse 'was

incremental, more or less keeping pace with tuff accumulation'. The fact that over 500 m of intracaldera tuffs accumulated subaqueously does not therefore imply water depths of over 500 m in the area of the caldera.

The misunderstanding over water depths may have arisen from our statement (p. 422) that depths of water equivalent to the supposed limiting depth for vesiculation of silicic magma (500 m) 'might be expected partially to suppress the eruption column and might explain the absence of any significant Plinian fall-out tuffs beneath the ash-flow deposits of the LRTF'. However, we pointed out that major subaerial ash-flow tuff eruptions may also lack a Plinian precursor phase and, as a result, its absence is not in itself diagnostic of a deep water eruption.

With the exception of a few details, such as the significance of hummocky cross-stratification in the intracaldera tuffs on Moel Hebog, which is irrelevant to the depth of emplacement of the primary ash-flow tuffs as it is the result of post-resurgence reworking, we are in broad agreement with Orton on the emplacement environments of the LRTF. His contention, however, that water depth at the onset of eruption 'would be shallower than that indicated by the underlying sandstone facies' is difficult to substantiate. Even from his log (Fig. 2) there is no indication of the 'fluidization or liquefaction of 3–5 m of the underlying sediment' which would destroy the evidence for shallowing and there is little indication of such disruption elsewhere. In fact the fining upward sequence (Howells *et al.* 1986, p. 412) is consistent with transgression prior to the main phase LRTF eruptions as substantiated by Orton (upper to lower shoreface-transitional offshore, Fig. 2).

The final part of Orton's contribution considers welding in subaqueous ash-flow tuffs and suggests that it is likely to be confined either to shallow water or to depths in excess of *c.* 500 m. He accepts that the extensively welded Yr Arddu Tuffs (Howells *et al.* in press), part of the initial stage of LRTF emplacement, were emplaced in water of intermediate depth, but designates them a 'special case' as they were ponded close to the vent. As the same may be said of the welded intra-caldera tuffs NW of the Gwynant Valley (p. 417), and as the welded facies at the base of the intracaldera tuffs in the SW were deposited subaerially, the discussion can be pursued little further with reference to the LRTF. Problems concerning the entry of hot pyroclastic flows into the sea, their persistence to intermediate depths (i.e. greater than the thickness of the deflated flow) and their capacity to weld are better addressed with reference to tuffs below the LRTF, such as the Pitts Head Tuff (A. J. Reedman *et al.* in preparation) or the Capel Curig Volcanic Formation (Francis & Howells 1973; Howells & Leveridge 1980; Howells *et al.* 1985) than in the context of the present paper.

References

Bourgeois, J. 1980. A transgressive shelf sequence exhibiting hummocky cross stratification: the Cape Sebastian Sandstone (Upper Cretaceous), SW Oregon. *Journal of Sedimentary Petrology*, **50,** 681–702.

Brenchley, P. J. & Cocks, L. R. M. 1982. Ecological associations in a regressive sequence: the latest Ordovician of the Oslo–Askee District, Norway. *Palaeontology*, **25,** 783–815.

Fernandez, H. E. 1969. Notes on the submarine ash flow tuff in Siargo Island, Surigao del Norte. *The Phillipine Geologist*, **23,** 29–36.

Fisher, R. V. 1984. Submarine volcaniclastic rocks. *In*: Kokelaar, B. P. & Howells, M. F. (eds) *Marginal Basin Geology*. Geological Society, London, Special publication **16,** 5–27.

Fiske, R. S. & Matsuda, T. 1964. Submarine equivelants of ash flows in the Tokiwa Formation, Japan. *American Journal of Science*, **262,** 76–106.

Francis, E. H. & Howells, M. F. 1973. Transgressive welded ash-flow tuffs among the Ordovician sediments of NE Snowdonia. *Journal of the Geological Society, London*, **129,** 621–41.

Howard, J. D. & Reineck, H. E. 1981. Depositional facies of high energy beach to offshore sequence: comparison with low energy sequence. *Bulletin of the American Association of Petroleum Geologists*, **65,** 807–30.

Howells, M. F. & Leveridge, B. E. 1980. *The Capel Curig Volcanic Formation*. Report, Institute of Geological Sciences, NO. 80/6.

——, Campbell, S. D. G. & Reedman, A. J. 1985. Isolated pods of subaqueous welded ash-flow tuff: a distal facies of the Capel Curig Volcanic Formation (Ordovician), North Wales. *Geological Magazine*, **122,** 175–80.

——, Reedman, A. J. & Campbell, S. D. G. 1986. The submarine eruption and emplacement of the Lower Rhyolitic Tuff Formation (Ordovician), N Wales. *Journal of the Geological Society, London*, **143,** 411–24.

——, Campbell, S. D. G., Reedman, A. J. & Tunnicliff, S. P. in press. A fissure-controlled acidic volcanic centre (Ordovician) at Yr Arddu, N Wales. *Geological Journal*.

Kokelaar, B. P. 1982. The fluidization of wet sediments during the emplacement and cooling of various igneous bodies. *Journal of the Geological Society, London*, **139,** 21–33.

——, Bevins, R. E. & Roach, R. A. 1985. Submarine silicic volcanism and associated sedimentary and tectonic processes, Ramsay Island, SW Wales. *Journal of the Geological Society, London*, **142,** 591–613.

Lipman, P. W. 1984. The roots of ash flow calderas in western North America: windows ino the tops of granitic batholiths. *Journal of Geophysical Research*, **89,** 8801–41.

Lockley, M. G. 1984. Faunas in a volcaniclastic flow of the Welsh basin: A synthesis of paleoecological and volcanological observations. *In*: Bruton, D. L. (ed.) *Aspects of the Ordovician System*. Palaeontological Contributions from the University of Oslo **295,** 195–201.

Moore, J. G. & Albee, W. C. 1981. Topographic and structural changes, March–July 1980, Photogrammetric data. *In*: Lipman, P. W. & Mullineaux, D. R. (eds) *The 1980 Eruptions of Mount St Helens, Washington*. U.S. Geological Survey Professional Paper 1250.

Pickerill, R. K. & Brenchley, P. J. 1979. Caradoc marine benthic communities of the South Berwyn Hills, North Wales. *Palaeontology*, **22,** 229–64.

Scherp, A. & Grabet, H. 1983. Unterdevonische Schmelzetuffe im rechtsrheinischen Schiefergebirge. *Neues Jahrbuch für Geologie und Palaeontologie*, **1,** 47–58.

Sparks, R. S. J., Sigurdsson, H. & Carey, S. N. 1980. The entrance of pyroclastic flows into the sea 2. Theoretical considerations on subaqueous emplacement and welding. *Journal of Volcanological and Geothermal Research*, **7,** 97–105.

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ENCLOSURE 3

Geol. Mag. **124** *(5), 1987, pp. 427–439. Printed in Great Britain*

The Pitts Head Tuff Formation: a subaerial to submarine welded ash-flow tuff of Ordovician age, North Wales

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Abstract – The Pitts Head Tuff Formation, of Ordovician (Caradoc) age, was emplaced as a thick (*c.* 700 m) intracaldera sequence and two outflow units comprising welded acidic ash-flow tuff. The Pitts Head pyroclastic flows were erupted subaerially but the lower and most extensive of the outflows crossed a shoreface, and continued for several kilometres offshore. The flow entered the sea without disruption and, following deflation and tuff emplacement, displaced the shoreface several kilometres to the east and northeast. Post-eruption subsidence in the northeast resulted here in the rapid establishment of environments deeper than had previously existed.

The lower outflow tuff is parataxitically to eutaxitically welded in both the subaerial and marine environments. The extremely regular plane-parallel welding foliation of the subaerial tuffs, however, contrasts with the locally highly deformed foliation of the tuff deposited beyond the shoreface. The deformed foliation, associated with irregular zones of intense siliceous nodule development, is ascribed to the upward streaming of water vapour generated at the tuff/sediment boundary. Elsewhere rheomorphism within the tuff was caused by instability resulting from emplacement on slopes related to faulting. Continued movement initiated extensive brecciation, detachment, and local gravity sliding of large rafts of tuff.

1. Introduction

The Pitts Head 'tuffs', initially regarded as extrusive rhyolites (Williams, 1927; Shackleton, 1959), were subsequently determined to be welded ash-flow tuffs or ignimbrites (Rast, Beavon & Fitch, 1958). Recent mapping in Snowdonia (British Geological Survey, 1985, 1986, in press) has covered most of the outcrop of these tuffs, now designated the Pitts Head Tuff Formation. It forms part of a thick and varied sequence of volcanics of Ordovician age erupted in a back-arc setting.

The formation has three main expressions which are geographically constrained. In the southwest, about Llwyd Mawr (Fig. 1B), *c.* 700 m of welded ash-flow tuffs are considered to represent an intracaldera fill (Roberts, 1969). On Moel Hebog the formation consists of two ash-flow tuffs (Fig. 2), each up to 90 m thick, both of which are interpreted as outflows from the caldera. This correlation has been supported geochemically (Howells *et al.* in press). North of Moel Hebog only the lower of these two outflows is present.

The formation is of Caradoc age, the lower outflow tuff lying approximately at the boundary between the Soudleyan and Longvillian stages. This tuff is generally underlain, and locally overlain, by various marine and non-marine sedimentary rocks which allow reconstruction of the palaeogeography immediately before and after its emplacement. In addition, the variable mesoscopic and microscopic characters of the outflow tuff can be related to subaerial and subaqueous emplacement environments and contemporaneous tectonism.

2. Sediments below the Pitts Head Tuff Formation: environmental interpretation

The sediments enveloping the Pitts Head Tuff Formation are the subject of a detailed analysis that is being prepared by one of us (G.O.) for publication elsewhere. Here the interpretations based on that analysis are presented, to establish the emplacement environments of the extensive lower outflow tuff of the Pitts Head Tuff Formation.

The base of the lower tuff is generally conformable with the underlying sediments. Only locally, at Moel Hebog, Moel Ddu and near Beddgelert (Fig. 1), where there has been post-emplacement sliding (see below) of large rafts of tuff, are the relationships discordant. The sediments are coarsest in the south, about Moel Hebog, and generally fine to the northeast. They reflect a transition, towards the northeast and east, from non-marine to shallow-marine conditions.

2.a. Southern area: Moel Hebog

Sediments below the lower Pitts Head Tuff at Moel Hebog (A, on Fig. 3A) comprise a prodelta sequence dominated by mudstone and siltstone, overlain by distributary mouth bar sediments and finally, by an alluvial plain and fan sequence immediately subjacent to the tuff.

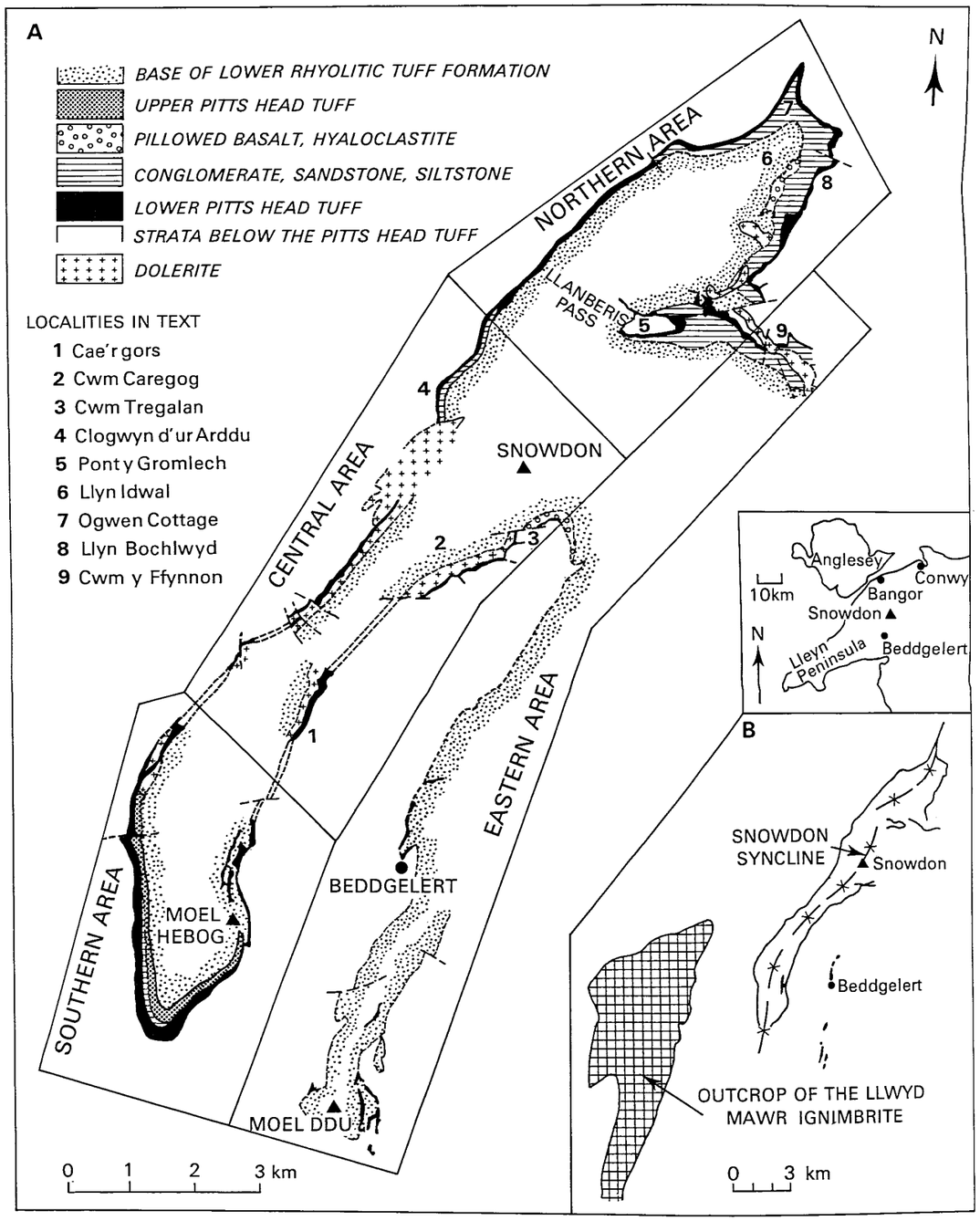

Figure 1. Maps showing the distribution of the Pitts Head Tuff Formation. (A) Distribution of the lower and upper Pitts Head Tuff outflow units, divided into areas described in the text. (B) Relationship between the outcrop of the outflow tuffs and of the intracaldera facies (Llwyd Mawr Ignimbrite).

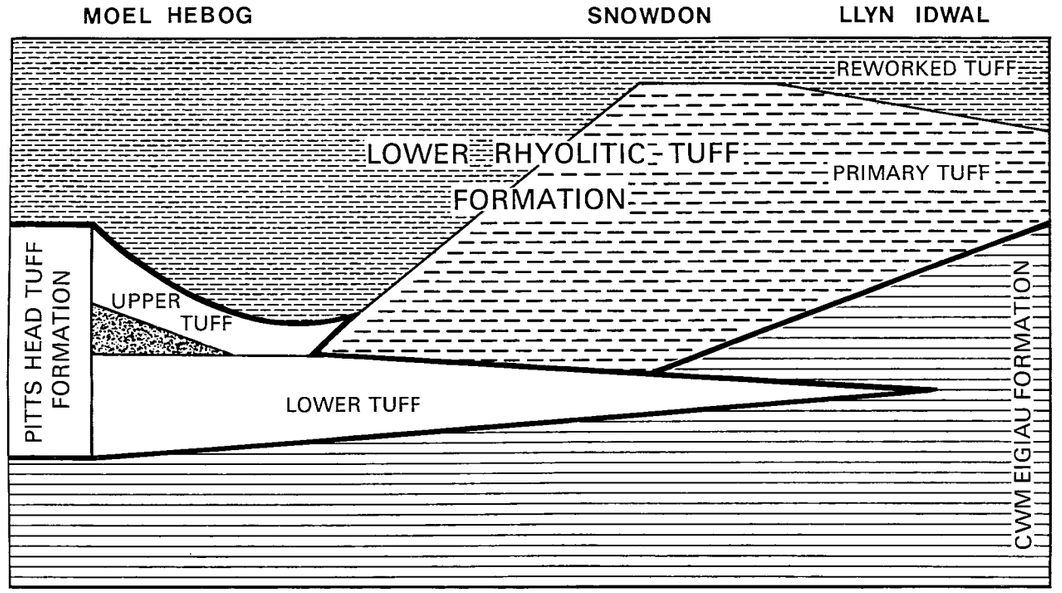

Figure 2. Vertical and lateral relationships of the Cwm Eigiau, Pitts Head Tuff and Lower Rhyolitic Tuff formations between Moel Hebog and Llyn Idwal. Dark stipple within the Pitts Head Tuff Formation represents sandstones and conglomerates.

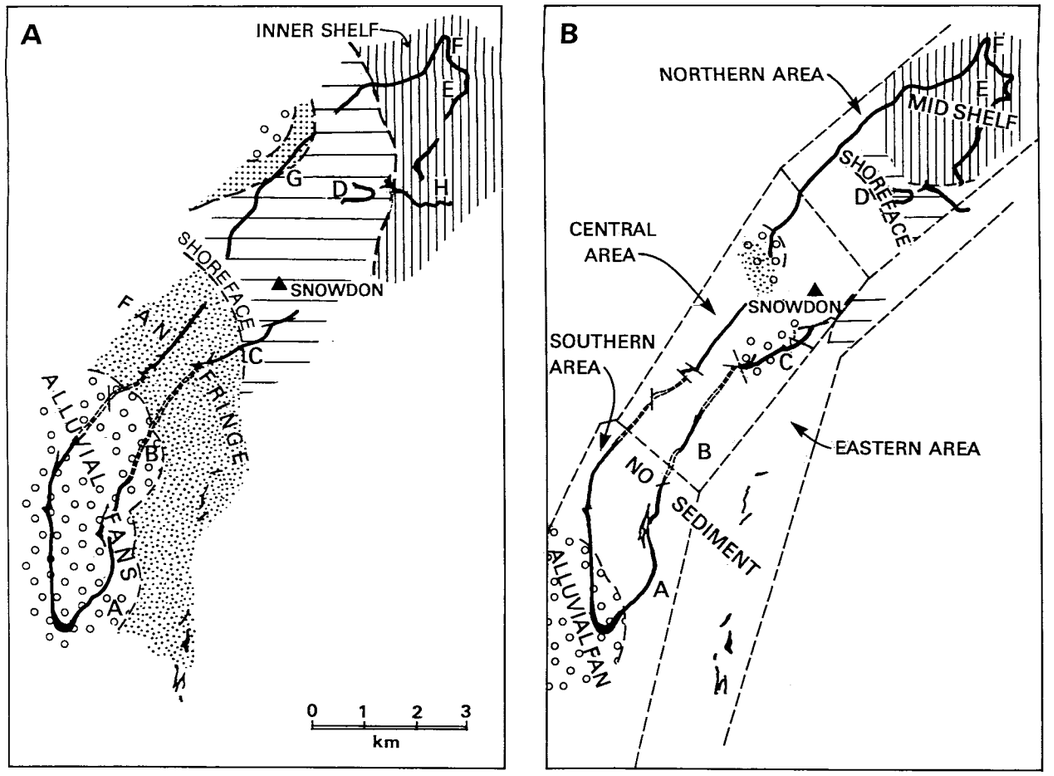

Figure 3. Emplacement environments of the lower Pitts Head Tuff. (A) Depositional environments of sediments immediately below the lower Pitts Head Tuff. Localities A to H are referred to in the text. (B) Depositional environments of sediments immediately overlying the lower Pitts Head Tuff. Areas outlined by pecked lines are those defined in Figure 1.

The alluvial plain sediments form tabular sheets which are dominated by trough cross-bedded, medium to coarse, pebbly sandstones typical of sandy braided stream deposits of mid to distal alluvial fans. A proximal-fan facies with evidence of channelling is also developed. In the lower part of the alluvial fan

sequence the absence of significant debris flows suggests that the fans had gentle slopes (Nilsen, 1982). Immediately subjacent to the lower tuff, however, the greater lateral variation in grain size and increased abundance of debris flow deposits indicates that smaller, steeper, or 'short-headed' fans (Flores, 1975) were developed as fan catchment areas were reduced prior to tuff emplacement. This was possibly in response to contemporaneous fault activity related to the onset of volcanism.

Within 10–40 cm of the base of the tuff, bedding structures in the underlying alluvial sediments have generally been destroyed. Where there has been post-emplacement mass movement and brecciation of the tuff (see below), the zone where bedding structures have been obliterated is up to 3 m thick.

2.b. Central area: Cae'r gors and Cwm Caregog

Sand-dominated sequences underlying the lower tuff in the central area are generally thinner than the equivalent sequences on Moel Hebog to the southwest. At Cae'r gors (B, Fig. 3A) on the eastern limb of the Snowdon Syncline, a 70 m thick sequence underlying the lower tuff is interpreted as having resulted from the northeastward progradation of alluvial fans. It comprises poorly sorted, massive, coarse-grained to conglomeratic sandstones deposited by debris flows and sheet floods. To the northeast, however, alluvial plain and distal alluvial fan deposits, the latter dominated by stacked sets of tangentially-based tabular cross-beds of medium-grained sandstone, are developed beneath the tuff. Further northeast, about Cwm Caregog (C, Fig. 3A), siltstones representing delta-front sediments coarsen upwards sharply to medium-grained sandstones containing abundant wave-swash reworked concentrates of detrital magnetite and ilmenite (cf. Reid & Frostick, 1985). These sandstones represent the progradation of wave-influenced beaches (Elliot, 1986)

2.c. Northern area: Llanberis Pass and Llyn Idwal

In Llanberis Pass the depositional environments of sediments immediately below the lower tuff are difficult to interpret because of relatively poor exposure and high tectonic strain. Similarly, below the tuff on the nearby Llechog Ridge structureless (?bioturbated), fine to very fine-grained sandstones lack positive evidence of their environment. To the north of the Llanberis Pass, however, at Gwastadnant (G, Fig. 3A), a clast-supported conglomerate up to 7 m thick, and containing rhyolite clasts up to 30 cm long, locally underlies the lower Pitts Head Tuff. The conglomerate is organized into 1–2 m thick, inversely- or non-graded beds and represents a short-headed fan delta dominated by debris flows and derived from the west and northwest. Finer sediments along strike to the north and south, as at Llechog, probably reflect shallow-marine conditions between fan salients.

At Llyn Cwm Ffynon (H, Fig. 3A), shallow-marine sediments are well exposed below the Pitts Head Tuff and comprise a coarsening upwards sequence of sandstones, commonly containing brachiopods in the bases of scours. Amalgamated sets of hummocky cross-stratification and low angle cross-bedding are common. Near the top of the sequence sandstones with large, straight-crested, symmetrical wave ripples pass up into coarse, swaley cross-stratified sandstones immediately below the Pitts Head Tuff. The sequence is interpreted as representing a prograding storm-dominated shoreline with lower shoreface sands being deposited immediately prior to the eruption of the Pitts Head Tuff.

Further north, about Llyn Idwal (E, F, Fig. 3A), the sandstones below the tuff contain brachiopod associations of the *Dinorthis–Macrocoelia* community suggesting water depths of less than 25 m (Pickerill, 1977; Pickerill & Brenchley, 1979). The sequence comprises shelf-ridge sands with interbar siltstones and mudstones. Fossiliferous, massive or cross-bedded, 5–20 cm thick sandstones within interbar siltstones reflect storm events (Brenchley, 1985).

3. The Pitts Head Tuff Formation

The Pitts Head ash-flow-tuffs are amongst the most distinctive in the Caradocian strata of North Wales. They are characterized by bleached weathered surfaces and a conspicuous welding foliation accentuated by silicification and by the presence of chloritized fiamme. The mesoscopic character of the extensive, 30–100 m thick lower tuff, changes systemically from south to north (Fig. 9) but outcrops of the upper tuff are limited to a small area in the south (Figs. 1, 2).

3.a. Southern area

About Moel Hebog (Fig. 1A), the lower tuff is up to 90 m thick with approximately 1 m of thinly bedded tuffs developed locally at its base (Fig. 4). The bedded tuffs are non-welded, vitroclastic and composed of identical material to the main flow unit and probably represent small precursor flows infilling subdued depressions in the pre-existing topography. The base of the main flow unit is non-welded and locally crystal-rich, grading rapidly up into welded tuff with distinctive chloritic fiamme below a conspicuous nodular zone (Fig. 4). The latter zone, 1–3 m thick and 3–5 m above the base of the tuff, is crowded with large siliceous nodules up to 40 cm in diameter (Fig. 5A). The welded tuff above the nodular zone is extremely uniform, with thin siliceous segregations, 1–5 mm thick, accentuating the planar, parataxitic, welding fabric (Fig. 6A). Locally, the top of the tuff contains siliceous nodules up to 5 cm.

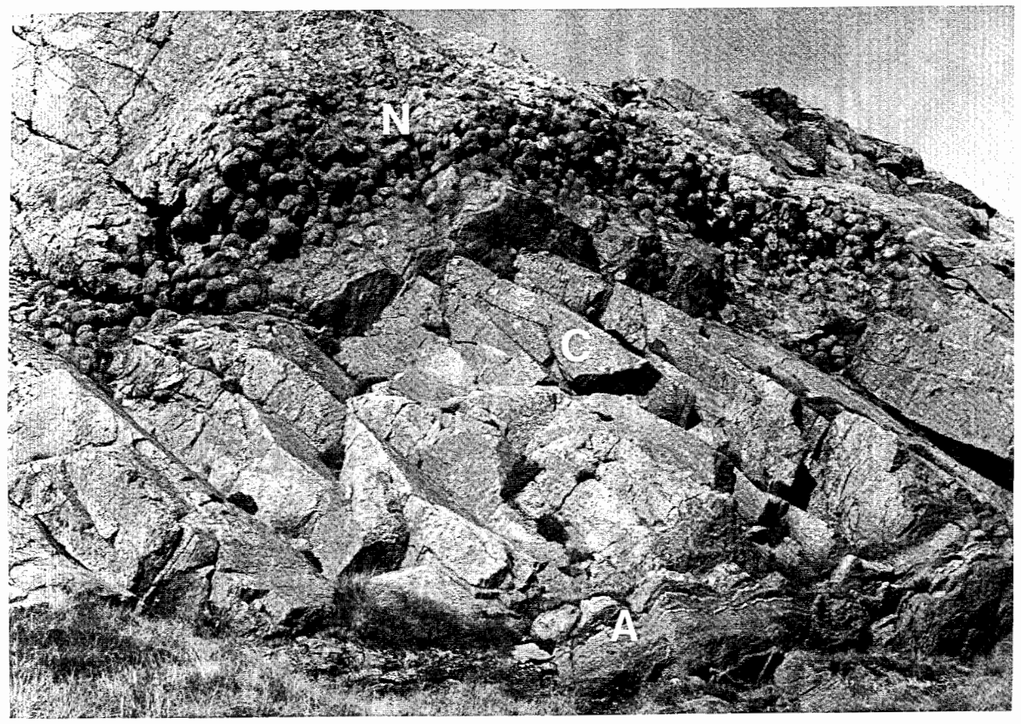

Figure 4. Section through the basal part of the subaerially emplaced lower Pitts Head Tuff on Moel Hebog. Basal bedded tuffs, A, are overlain by non-welded to weakly welded tuff passing up into welded tuff with crudely developed columnar joints, C, and zone of siliceous nodules, N.

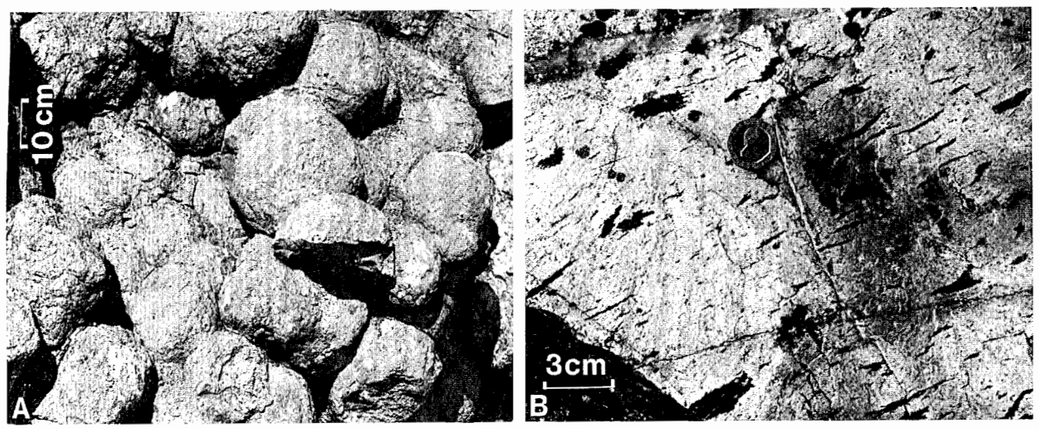

Figure 5. (A) Large siliceous nodules developed in the basal nodular zone (see Fig. 4) of the lower Pitts Head Tuff, Moel Hebog. (B) Chloritized fiamme in the lower Pitts Head Tuff, Ogwen Cottage.

In places on the east and northeast flanks of Moel Hebog the tuff is distinctively brecciated. The breccias comprise rotated angular clasts of welded tuff (Fig. 6A, B) occurring either in thin discontinuous zones, commonly aligned along joints, or as major zones which in extreme cases occupy the entire thickness of the tuff unit. Sections exposed on the eastern slopes of Moel Hebog display systematic changes in the lowest few metres of the tuff as areas of pervasive brecciation are approached (Fig. 7A). Traced laterally towards the brecciated tuff, the crystal-rich, weakly or non-welded, basal zone locally thins, develops a lobate downward protrusion into the underlying sediments and is entirely absent where the flow is brecciated

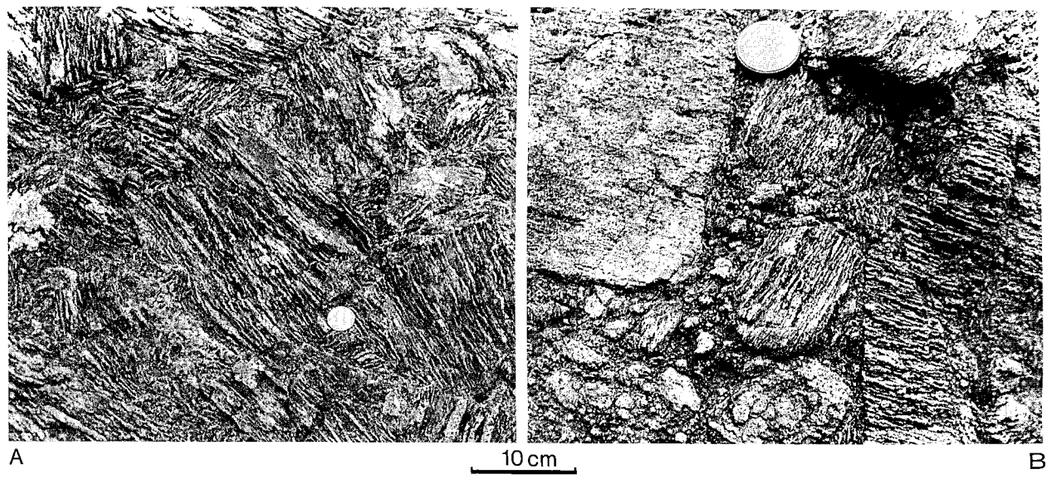

Figure 6. (A) Brecciated lower Pitts Head Tuff, Moel Hebog, with extremely angular clasts. Siliceous ribs accentuate the welding foliation within the clasts. (B) Brecciated lower Pitts Head Tuff, Moel Hebog, displaying abraded, sub-angular to sub-rounded clasts.

throughout. Similarly the zone of large siliceous nodules becomes diffuse and is not developed within the brecciated tuff. Large-scale brecciation occurs abruptly across a zone of vertical fracturing, individual clasts are up to 3 m in length and many of the largest blocks are aligned with their welding foliation approximately normal to the base of the flow. The base of the breccia is discordant with the underlying sediments. Locally the tuff is totally disrupted with part of the breccia detached from the parent body and the intervening space infilled with ash-flow tuff of the overlying Lower Rhyolitic Tuff Formation containing clasts of the Pitts Head Tuff (Fig. 7A).

On the southwest flanks of Moel Hebog, the upper Pitts Head Tuff, up to 70 m thick, differs from the lower only in its lack of a basal nodular zone. The tuff wedges out abruptly in exposures southeast of the summit and does not extend to the northeast of Moel Hebog (Fig. 1A), where only the lower tuff is present.

3.b. Central area

In the central area the lower tuff, 70–100 m thick, is parataxitically welded throughout and, in contrast to its southern crop, the basal non-welded tuff and nodular zone are generally absent.

3.c. Northern area

In the northern area the lower tuff is characteristically parataxitically to eutaxitically welded and contains conspicuous fiamme of chloritized collapsed pumice (Fig. 5B). At Pont-y-Gromlech (5, Fig. 1A) and to the north of the Llanberis Pass the eutaxitic welding foliation is locally highly irregular and discordant with the upper and lower contacts of the tuff. In places the foliation forms well-defined mesoscopic folds with brecciated cores, and Wright & Coward (1977) have described diapir-like structures, up to 30 m in diameter, defined by the welding foliation and associated with intense nodule development. The siliceous nodules typically occur in irregular zones scattered through the body of the tuff and locally in a more persistent zone near its top. On the south side of the Llanberis Pass, the top of the tuff is mantled by up to 1.5 m of non-welded vitric dust-tuff.

In the most northerly outcrops, about Llyn Idwal (6, Fig 1A) the lower tuff varies from 30–50 m thick. At its base 1–2 m of dark grey, cleaved, feldspar-rich, non-welded tuff grade up into 25–45 of white weathered eutaxitically welded tuff with dark chloritic fiamme (Fig. 5B). Columnar jointing is locally developed. Siliceous segregations accentuate the welding foliation but are less well developed than in the south. The orientation of the welding foliation is locally highly variable indicating rheomorphic deformation. Zones packed with siliceous nodules, 1–3 cm in diameter, occur irregularly through the body of the tuff.

To the west of Llyn Idwal, where the tuff thickens across a contemporaneous fault, rafts of bedded sandstone, up to 4 m thick, occur towards its top. East of Gribin Ridge, near Llyn Bochlwyd (8, Fig. 1A), up to 30 m of vertical movement on a growth fault caused substantial thickening of the tuff within a half-graben, and subsequent movement at the base of the tuff resulted in the intrusion of sandstone dykes up to 10 m wide.

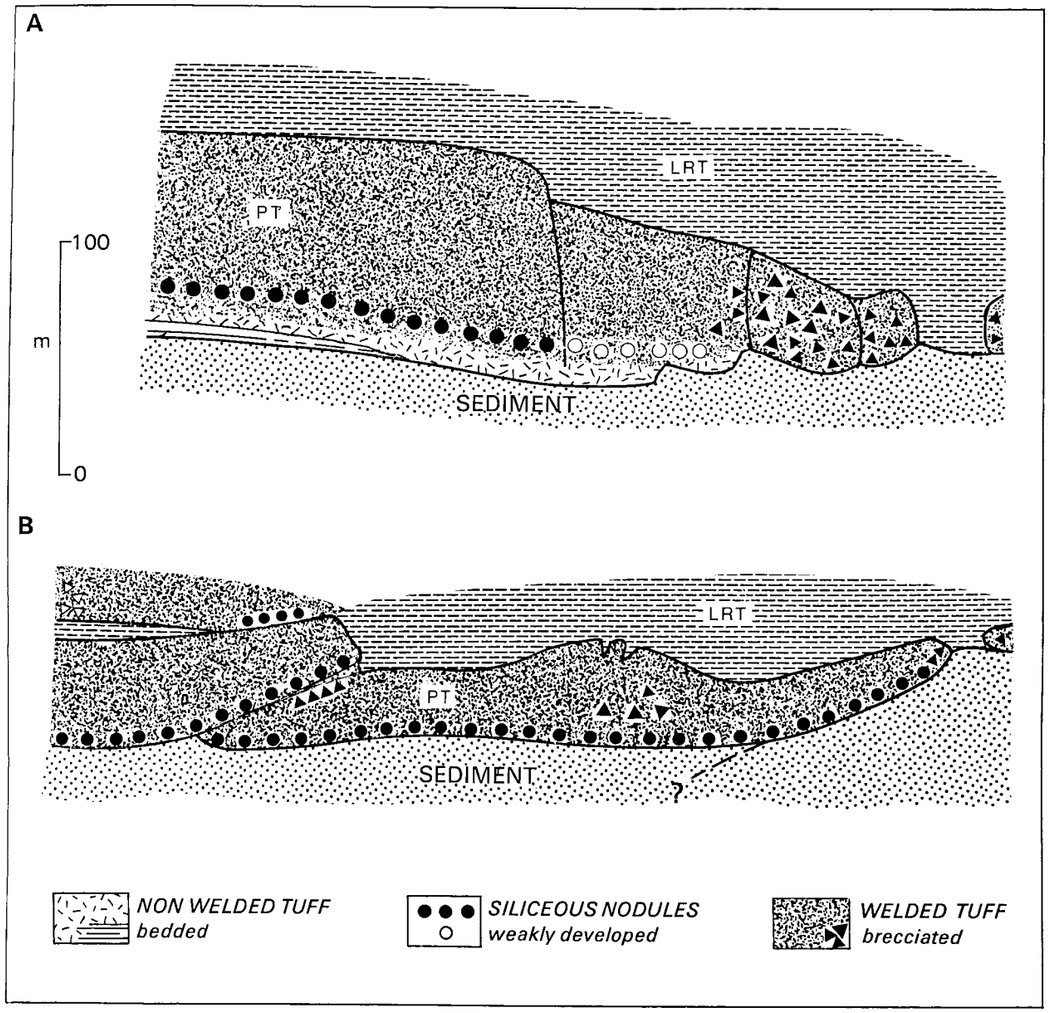

Figure 7. (A) Cross-section through the lower Pitts Head Tuff, northeast of the summit of Moel Hebog, displaying lateral lithological variations in the tuff. PT = Pitts Head Tuff; LRT = Lower Rhyolitic Tuff Formation. (B) Cross-section through the lower Pitts Head Tuff at Moel Ddu showing rheomorphic deformation and stacking of rafts of tuff. Vertical scale as in part (A). Thin non-welded zone at base of tuff not shown. PT = Pitts Head Tuff; LRT = Lower Rhyolitic Tuff Formation.

3.d. Eastern area

East of the Snowdon Syncline (Fig. 1B) the distribution of the Pitts Head Tuff is limited and its character changes. East of the Llanberis Pass (Fig. 1A) the tuff progressively thins over a distance of *c.* 3 km, contains increasing amounts of epiclastic material, and at Cwm y Ffynnon (9, Fig. 1A) is represented by thin beds of reworked tuff interbedded with sediments. Similarly the tuff changes across Cwm Tregalan (3, Fig. 1A); the locally brecciated tuff on the west side of the Cwm is represented on the east side by several thin beds of coarse sandstone composed entirely of angular clasts of welded tuff.

In the southeast, near Beddgelert and Moel Ddu, the tuff occurs as large allochthonous rafts and blocks, some several hundred metres long, either stacked upon each other or isolated at different stratigraphic levels within the post-lower Pitts Head Tuff sequence. Most of the allochthonous tuff rafts are either partially or completely brecciated and some display welding foliations oriented at a high angle to their lower and upper contacts.

The lowest and most easterly exposure of the tuff at

Moel Ddu, 30 m thick, is mainly autochthonous. It displays characters similar to the lower tuff at Moel Hebog: a non-welded to weakly welded basal zone, 1–3 m thick, overlain in turn by a zone of large siliceous nodules 2–3 m thick and about 25 m of intensely welded tuff displaying a plane-parallel, concordant welding foliation accentuated by silicification. In the vicinity, various features related to local mass movement of the tuff are also seen. Near the northern edge of the outcrop, the basal contact of the tuff becomes discordant with the underlying sandstones, and the welding foliation, steeper in the basal zone than the contact, steepens further through vertical to 'overturned' at the top of the unit. Further to the north, the body of the tuff is disrupted and a large raft is detached from the main crop (Fig. 7B). Similar post-emplacement mass movement has been determined in the central part of the outcrop where the tuff is locally repeated in the sequence by vertical stacking (Fig. 7B). In these tuffs, a variably developed lineation within the welding foliation is defined by elongation of siliceous lenticles and by the long axes of elliptical siliceous nodules. A survey of the eastern outcrops indicates that the lineation trends between north-northwest and northeast and plunges at 0–30°, either to the north or south. The lineation postdated, or was coincident with, nodule growth but predated brecciation.

Brecciation of the tuff is common about Moel Ddu and Beddgelert. In the autochthonous tuff at Moel Ddu, breccia, comprising angular clasts up to 10 cm, occurs as irregular zones and thin sheets along joints. Locally, conjugate sets of kink bands deform the welding fabric (Fig. 8A). and extend laterally into fractures which define incipient clast boundaries (Fig. 8B). Small breccia dykes, 20–50 cm wide, containing sub-rounded clasts, probably resulted from volatile-streaming along fractures.

Most of the large rafts of Pitts Head Tuff within the Lower Rhyolitic Tuff Formation at Moel Ddu consist either entirely of breccia or have a welding foliation oriented at a high angle to their basal contacts. They all lack the typical basal non-welded zone and nodule zone.

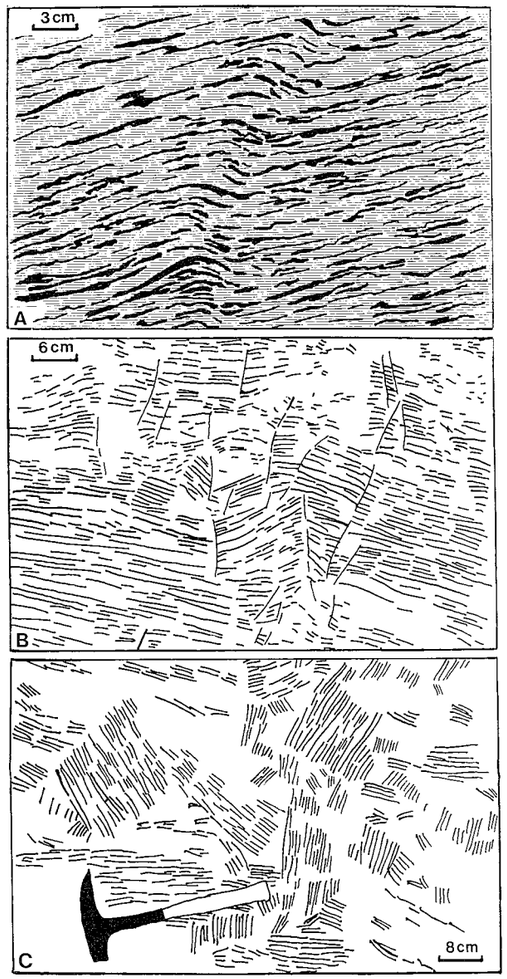

Figure 8. Stages in tuff brecciation traced from field photographs. (A) Part of a kink band defined by deformed welding foliation. (B) Anastomosing fractures defining incipient clast boundaries. (C) Rotated angular clasts in completely brecciated tuff.

3.e. Microscopic character

Typically, the welded tuffs are devitrified and recrystallized and comprise a fine aggregate of sericite, quartz, feldspar and chlorite in variable proportions, with impersistent ribs and lenses of coarser recrystallized quartz accentuating the welding fabric. Shards are locally well defined, although generally they have been overprinted by recrystallization. Euhedral, subhedral and fragmented phenocrysts of albite–oligoclase feldspar (< 2 mm), variably altered, are particularly concentrated in the basal zone of the tuff. Above, they are isolated and rarely exceed 5% of the total volume. In places the phenocrysts display imbrication but normally show no preferred orientation. Lithic fragments are mainly of siltstone or fine sandstone, some with irregular indented margins indicative of incorporation into the flow when only partially lithified. They occur mainly in the basal zone in the northern part of the outcrop. Perlitic fracturing of the matrix is common in the north, occurring in the central part of the flow, but on the southern flanks of Snowdon is restricted to the lowest few metres, indicating welding to the base.

In thin-section siliceous nodules as small as 0.04 mm diameter are seen. They comprise a quartz mosaic, and locally the welding foliation can be distinguished

partially passing into, and partially deflected around the nodules, indicating that they developed during compactional welding.

Non-welded vitroclastic textures are present only in the bedded tuffs beneath the lower tuff at Moel Hebog, in the basal, crystal-rich tuff near Moel Hebog, and at Ogwen Cottage (7, Fig. 1A), and in the vitric dust-rich top of the tuff north of Snowdon.

4. Sediments above the Pitts Head Tuff Formation

4.a. Southern area: Moel Hebog

On the southeast side of Moel Hebog (A, Fig. 3B) the lower Pitts Head Tuff is directly overlain by the upper tuff. However, on the west and southwest sides of Moel Hebog a sequence of medium to coarse-grained sandstone overlain by a boulder conglomerate, which thickens southwestwards to 40 m, separates the two tuffs. The conglomerate is massive, with a framework of well rounded sandstone, siltstone and acid volcanic clasts, up to 30 cm in diameter, and a sparse silt and sand matrix. It is interpreted as a proximal alluvial fan which thins rapidly to the northeast and is overstepped by the upper tuff. On the northwest and northeast slopes of Moel Hebog, where the upper tuff is absent, the lower tuff is directly overlain by primary ash-flow tuffs of the Lower Rhyolitic Tuff Formation (Howells, Reedman & Campbell, 1986).

4.b. Central area: Cae'r gors and Cwm Tregalan

Sediments are discontinuously and at best thinly developed above the lower tuff in the central area. In the south, near Cae'r gors (B, Fig. 3B), the tuff is overlain by ash-flow tuffs of the Lower Rhyolitic Tuff Formation with a thick dolerite sill intruded along the contact. Locally, south of Cwm Caregog, a thin, medium to coarse-grained, pebbly sandstone lies between the dolerite and the upper surface of the tuff. To the northeast, into Cwm Tregalan (C, Fig. 3B), the dolerite is interlayered with thin pebbly sandstones. On the north and east sides of Cwm Tregalan a thick sequence of extrusive submarine basaltic pillow breccias occurs above beds of coarse sandstone almost entirely composed of clasts of the lower Pitts Head Tuff.

4.c. Northern area: Pont-y-Gromlech and Llyn Bochlwyd

Sediments overlying the tuff are well exposed north of Snowdon. At Pont-y-Gromlech (D, Fig. 3B), the fine-grained top of the tuff is overlain by a fining upward, sandstone-dominated sequence containing brachiopods characteristic of the *Dinorthis–Macrocoelia* community. Sedimentary structures include trough cross-bedding, swaley cross-stratification and low-angle cross-lamination to horizontal lamination. Above, the strata are characterized by planar bedding, thick (*c.* 20 cm) winnowed fossiliferous beds, coquina-filled scours and occasional low to moderately inclined cross-bedding with shells along the foresets. The sequence is interpreted as representing a transgressive non-barred, wave-influenced shoreline.

To the northeast, at Cwm Cneifion, south of Llyn Bochlwyd (E, Fig. 3B), the tuff is overlain by *c.* 10 m of siliceous siltstone in turn overlain by sandstones. Transported shells of the *Dinorthis* community occur either as basal lags to hummocky cross-stratification or within massive graded beds. The facies are typical of mid to outer storm-dominated shelves (Brenchley, 1985).

5. Discussion

The Pitts Head Tuff Formation forms a virtually continuous outcrop for *c.* 20 km from Moel Hebog, southwest of Snowdon to the vicinity of Ogwen Cottage in the north. On Moel Hebog the formation comprises two ash-flow tuffs, averaging 80 and 50 m in thickness, but to the north is represented solely by the lowest tuff which lies approximately at the Soudleyan-Longvillian stage boundary. The lithological similarity of these tuffs to the *c.* 700 m thick Llwyd Mawr ignimbrite southwest of Moel Hebog (Roberts, 1969) has long been recognized, though originally both were thought to be extrusive rhyolites (Shackleton, 1959). Roberts (1969) considered that the Llwyd Mawr ignimbrite, with its associated rhyolites, represented a caldera infill and was the source of the outflow Pitts Head tuffs. Studies of vertical variations in the intensity of the welding and deformation of pumice (Roberts, 1969; Roberts & Siddans, 1971) through the Llwyd Mawr sequence indicated the presence of two major flow units, the lower *c.* 70 m and the upper > 500 m thick, which comprised a single cooling unit.

The form of the original eruptive centre is difficult to define as the Llwyd Mawr ignimbrite is folded and is either in faulted contact with, or overlies, graptolitic mudstones which are considerably older (Llanvirn). It displays no preserved relationships with younger strata. It is probable that the great thickness of tuffs, in particular the upper flow unit, accumulated during incremental caldera collapse, the elongate outcrop suggesting that the form of the 'caldera' was a NNE-trending graben or half-graben. It is not known whether the lower and upper Pitts Head Tuff of Moel Hebog correlate respectively with the lower and upper units of the Llwyd Mawr ignimbrite. However, the evidence of a considerable time interval between the eruption of the lower and upper Pitts Head Tuff suggests that this correlation is unlikely. In the northern area (Fig. 1) up to 200 m of marine sediments were deposited in the interval between the eruption of lower and upper Pitts Head Tuff, indicating that they are the products of two distinct eruptive cycles.

Sandstones immediately below the lower Pitts Head Tuff on Moel Hebog represent the deposits of alluvial fans, and the sandstones and conglomerates above the lower tuff represent proximal-fan sediments. Both are derived from the west and southwest and contain abundant acid tuff and rhyolitic debris. This indicates the local extrusion, possible uplift and erosion of acid volcanics in the vicinity of the Llwyd Mawr centre prior to caldera collapse and the emplacement of the lower Pitts Head Tuff. This in turn was followed by resurgence and further subaerial erosion of the caldera complex prior to the emplacement of the upper Pitts Head Tuff.

From the evidence of the substrate, the lower Pitts Head outflow tuff was deposited subaerially on an east to northeast sloping terrain from Moel Hebog in the south to the vicinity of Cwm Caregog in the north (Fig. 3A), where the ash flow crossed a shoreline. There is no evidence of disruption of the flow at the shoreline, and the flow continued for at least 10 km into the sea, while retaining sufficient heat for most of the tuff to weld following emplacement. Clearly the hot gas-charged flow did not ingest any substantial quantity of water on entering the sea. Sparks, Sigurdsson & Carey (1980) have modelled the thermal behaviour of hot pyroclastic flows under water, indicating that subaqueous welding is theoretically possible. Howells, Campbell & Reedman (1985) have suggested that hot subaqueous pyroclastic flows might develop a carapace of steam, generated by film boiling (the Leidenfrost effect) at the flow/water interface, which would both facilitate the passage of the flow through water and provide insulation. The upper, less dense, part of the flow would not enter the sea but travel across the surface, deflate and form a fine-grained, water-settled ash deposit above the subaqueous flow deposit. This may be represented by the non-welded vitric dust-tuff locally present at the top of the lower Pitts Head Tuff in its northernmost outcrops, though, on the basis of the sedimentology and palaeoecology of the enveloping sediments, it remains debatable as to whether water depths were great enough for the tuff to be entirely submerged upon emplacement even at its most distal end.

The brachiopod communities in the marine sediments below the most distal outcrops of the tuff suggest that water depths did not exceed a few tens of metres, indicating a submarine palaeoscope of *c.* 2°. The thickness of the lower tuff varies from 80 to 35 m north of Snowdon and its thickness immediately after emplacement, prior to compaction and welding, would have been significantly greater requiring substantial displacement of the shoreline. A transgressive shoreface facies has been determined immediately above the tuff in Llanberis Pass, approximately 5 km north of the postulated pre-tuff shoreline in the vicinity of Cwm Caregog and 2 km east of that at Gwastadnant. A further 3–5 km to the north, in Cwm Cneifion and near Llyn Idwal (E, F, Fig. 3B), the sediments above the tuff display facies typical of a storm-dominated outer shelf, a deeper water environment than that seen immediately below the tuff. Such a change implies rapid sea floor subsidence during or immediately after tuff emplacement. Maximum subsidence occurred in the north, about Llyn Idwal (6, Fig. 1A), where over 200 m of marine sediments were deposited between the top of the lower Pitts Head Tuff and the base of the Lower Rhyolitic Tuff Formation, as compared to only 2 m of pebbly sandstone at Clogwyn d'ur Arddu (4, Fig. 1A).

The lower tuff is restricted to the west side of Snowdonia. In the north and central parts of the eastern area, the primary ash-flow tuff passes abruptly eastwards into reworked tuffs interbedded with marine sandstones, as at Cwm y Ffynnon and Cwm Tregalan (Fig. 1). In the southeast the lower tuff is present mainly as allochthonous rafts and blocks, but a large remnant of the autochthonous tuff sheet indicates that it extended as far east as Moel Ddu. About Yr Arddu, 3 km further to the east (Howells *et al.* in press), the Pitts Head Tuff ash flows are absent, though the Pitts Head eruptions may be locally represented by a water-settled accretionary lapilli tuff intercalated in the marine sandstones. In the eastern area a NE-trending fault zone, extending from the vicinity of Moel Ddu through Beddgelert (Fig. 1), was active immediately before, during and after the emplacement of both the lower Pitts Head Tuff and the ash-flow tuffs of the Lower Rhyolitic Tuff Formation (Rast, 1961; Beavon, 1980; Howells, Reedman & Campbell, 1986). Early movement on these faults may have produced a tilt-block topography which restricted the distribution of the lower Pitts Head ash-flow tuff to the western side.

From southwest to northeast, the change of emplacement environment of the lower Pitts Head tuff, from subaerial to subaqueous, and proximal to distal, is reflected in post-emplacement modifications of the tuff fabric (Fig. 9). In the south, close to source, the tuff was emplaced on alluvial fans and fan fringes and is characterized by a non-welded, locally bedded base, a conspicuously developed nodular zone above the non-welded zone, a concordant planar parataxitic welding fabric accentuated by silica segregation through the rest of the flow, and an eroded top. Further north, in the vicinity of the shoreline, the tuff is welded to the base and lacks the conspicuous nodular zone but otherwise preserves the characters seen in the south. In the subaqueous environment the tuff is characterized by a crystal-rich, non or weakly-welded base, a thick welded zone with distinct chloritic fiamme and a eutaxitic fabric, commonly rheomorphically folded and associated with irregular zones of intense siliceous nodule development and a non-welded top of vitric dust-tuff locally displaying evidence of marine reworking. The gradual change from a parataxitic, intensely silicified fabric in the south to a eutaxitic and more sporadically silicified fabric in the north reflects mainly a proximal to distal decline in emplacement temperature. Similarly, except near the base of the flow, collapsed pumice clasts have generally been totally replaced during silicification in the proximal tuff but are preserved in the distal tuff as chloritized fiamme. However, variations in the style of siliceous nodule development, the degree of welding at the base of the flow, the plastic deformation of the welding fabric, and erosion at the top of the flow are related to contrasting emplacement environments rather than distance from the eruptive source.

The growth of large, 10–40 cm, siliceous nodules at a distinct horizon above the non-welded base occurred where the tuff was deposited upon alluvial fans and fan fringe sediments, and resulted from the entrapment of volatiles near the base of the zone of most intense welding. The relationship of these nodules to the welding foliation shows that their growth took place during the later stages of compactional welding but prior to its completion. Locally, as at Moel Ddu, the nodules were either stretched or continued to grow during downslope rheomorphic flow.

Welding to the basal contact, at or immediately beyond the shoreline, may have been aided by the absorption of water vapour at the base of the flow which facilitated pressure solution at shard contacts. This early welding at the base reduced porosity and inhibited the upward migration of volatiles and the growth of large nodules. Further beyond the shoreline, however, a non-welded crystal-rich base is again developed, perhaps reflecting the limited but progressive absorption of water at the base of the flow, and consequent cooling, during transport. Also in the inner shelf environment, rheomorphism resulted in ductile deformation of the welding foliation and the formation of crude diapir-like structures associated with intense nodule development (Fig. 9). Wright & Coward (1977) have ascribed the structures to the upward streaming of volatiles, mainly water vapour, from the base of the tuff, and consequently these nodules may represent original gas cavities, lithophysae, later infilled by silica.

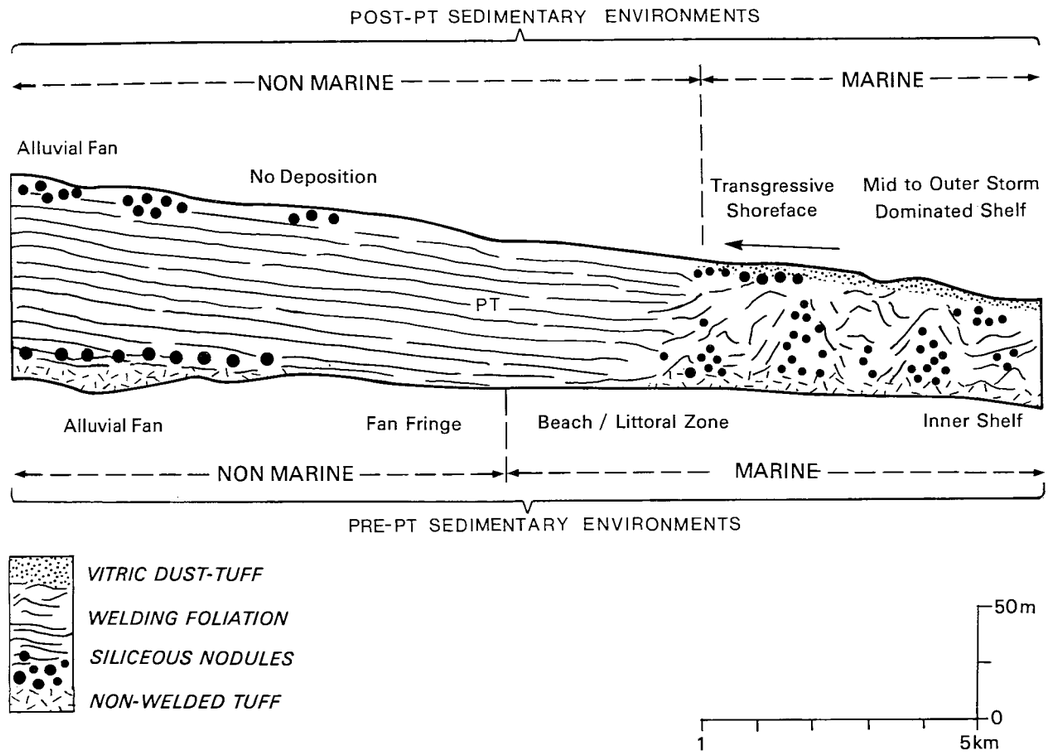

Figure 9. Schematic cross-section of the lower Pitts Head Tuff showing the major lateral variations in tuff lithology and the subjacent and suprajacent sedimentary environments.

In the submarine environment a non-welded vitric dust-tuff top to the flow unit was locally preserved and variably reworked, but in the subaerial environment the non-welded top was removed prior to the deposition of the suprajacent strata (Fig. 9).

Post-emplacement mass gravity flowage and brecciation of the tuff was related to contemporaneous tectonism in the eastern area, east of Moel Hebog, and about Moel Ddu and Beddgelert (Fig. 1A). The ramped welding fabric observed at Moel Ddu, and the local presence of a stretching lineation that predated brittle fracture and breccia formation, indicates that movement occurred while the tuff was still plastic. The lateral movement was very small and open crevasses at the top of the tuff (Fig. 7B), subsequently infilled by later deposits, indicate a rapid transition to brittle behaviour. In the body of the flow the transition from ductile to brittle deformation is represented by conjugate sets of kink bands. Subsequent extensive brittle fracture resulted in the pervasively brecciated tuff. Breccia clasts were locally mobilized by volatile streaming to form breccia dykes. Where the tuff is most pervasively brecciated its base becomes discordant, and the orientation of the largest clasts, at a high angle to the base, indicates that they have been rotated during mass movement. Both early viscous flow and subsequent brecciation resulted from internal instability, probably caused by change in the configuration of the substrate as a result of contemporaneous faulting (Fig. 10A).

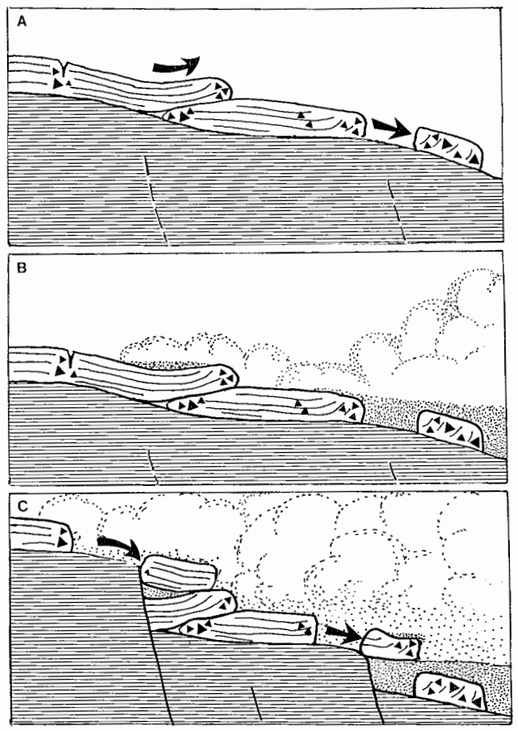

Figure 10. Cartoon illustrating post-emplacement mass movement of the Pitts Head Tuff on slopes produced by penecontemporaenous faulting. (A) Rheomorphic flow, brecciation and ramping of the tuff prior to the eruption of the Lower Rhyolitic Tuff Formation. Subjacent sediments only partially lithified. (B) Eruption of early pyroclastic flows of the LRTF. (C) Downslope collapse and stacking of rafts of Pitts Head Tuff during fault reactivation within and at the southwest margin of the LRTF caldera, coincidental with the eruption of the main phase of the LRTF.

The restriction of the upper Pitts Head outflow tuff to the area about Moel Hebog, and the downslope collapse of the lower tuff in the same area, were influenced by the development of the Lower Rhyolitic Tuff Formation (LRTF) caldera (Howells, Reedman & Campbell, 1986). The northeastern limit of the upper tuff coincides precisely with the southwestern limit of the LRTF primary ash-flow tuff sheet at the southwestern margin of the LRTF caldera (Fig. 2). Eruption of the LRTF ash flows modified the pre-existing palaeogeography; most significantly, the easterly to northeasterly palaeoslope, upon which the lower Pitts Head Tuff and the superjacent alluvial fan had been deposited, was reversed, thus restricting the northward flow of the upper Pitts Head Tuff. The reactivation of faults with downthrows to the north-east, at the southwestern margin of the LRTF caldera, caused the collapse and lateral movement of large rafts of the Pitts Head tuffs into the accumulating LRTF sequence (Fig. 10B, C).

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References

Beavon, R. V. 1980. A resurgent cauldron in the early Palaeozoic of Wales, U.K. *Journal of Volcanology and Geothermal Research* **7**, 157–74.

Brenchley, P. J. 1985. Storm influenced sandstone beds. *Modern Geology* **9**, 369–96.

British Geological Survey. 1985. 1:50000 *Sheet* 106 (*England & Wales*), *Bangor*. Southampton: Ordnance Survey.

British Geological Survey. 1986. 1:25000 *Sheet SH* 65 *N*/66 *S*, *Passes of Llanberis & Nant Ffrancon*. Southampton: Ordnance Survey.

British Geological Survey. In press. 1:25000 *Sheet SH* 64 *N*/65 *S*, *Snowdon*. Southampton: Ordnance Survey.

Elliot, T. 1986. Siliclastic shorelines. In *Sedimentary Environments and Facies* (ed. H. G. Reading), pp. 155–88. Oxford: Blackwells.

Flores, R. M. 1975. Shortheaded stream deltas: Model for Pennsylvanian Haymond Formation. *Bulletin of the American Association of Petroleum Geologists* **59**. 2288–2301.

Howells, M. F., Campbell, S. D. G. & Reedman, A. J. 1985. Isolated pods of subaqueous welded ash-flow tuff: a distal facies of the Capel Curig Volcanic Formation (Ordovician), North Wales. *Geological Magazine* **122**, 175–80.

Howells, M. F., Reedman, A. J. & Campbell, S. D. G. 1986. The submarine eruption and emplacement of the Lower Rhyolitic Tuff Formation (Ordovician), N. Wales. *Journal of the Geological Society of London* **143**, 411–23.

Howells, M. F., Campbell, S. D. G., Reedman, A. J. & Tunnicliff, S. P. In press. A fissure-controlled acidic volcanic centre (Ordovician) at Yr Arddu, N. Wales. *Geological Journal*.

Nilsen, T. H. 1982. Alluvial fan deposits. In *Sandstone Depositional Environments* (eds. P. A. Scholle and D. Spearing), pp. 49 86. American Association of Petroleum Geologists Memoir no. 31.

Pickerill, R. K. 1977. Trace fossils from the Upper Ordovician (Caradoc) of the Berwyn Hills, Central Wales. *Geological Journal* **12**, 1–16.

Pickerill, R. K. & Brenchley, P. J. 1979. Caradoc marine benthic communities of the south Berwyn Hills, North Wales. *Palaeontology* **22**, 229–64.

Rast, N. 1961. Mid-Ordovician structures in south-western Snowdonia. *Liverpool and Manchester Geological Journal* **2**, 645–52.

Rast, N., Beavon, R. V. & Fitch, F. J. 1958. Sub-aerial volcanicity in Snowdonia. *Nature* **181**, 508.

Reid, F. & Frostick, L. E. 1985. Beach orientation, bar morphology and the concentration of metalliferous placer deposits: a case study, Lake Turkana, N. Kenya. *Journal of the Geological Society of London* **192**, 837–48.

Roberts, B. 1969. The Llwyd Mawr Ignimbrite and its associated volcanic rocks. In *The Pre-Cambrian and Lower Palaeozoic rocks of Wales* (ed. A. Wood), pp. 337–56. Cardiff: University of Wales Press

Roberts, B. & Siddans, A. W. B. 1971. Fabric studies in the Llwyd Mawr Ignimbrite, Caernarvonshire, North Wales. *Tectonophysics* **12**, 283–306.

Shackleton, R. M. 1959. The stratigraphy of the Moel Hebog district between Snowdon and Tremadoc. *Liverpool and Manchester Geological Journal* **2**, 216–52.

Sparks, R. S. J., Sigurdsson, H. & Carey, J. N. 1980. The entrance of hot pyroclastic flows into the sea, II. Theoretical considerations on subaqueous emplacement and welding. *Journal of Volcanology and Geothermal Research* **7**, 97–105.

Williams, H. 1927. The Geology of Snowdon (North Wales). *Quarterly Journal of the Geological Society of London* **83**, 346–431.

Wright, J. V. & Coward, M. P. 1977. Rootless vents in welded ash-flow tuffs from northern Snowdonia, N. Wales, indicating deposition environment of shallow water. *Geological Magazine* **114**, 133–40.

ENCLOSURE 4

Fan Deltas: Sedimentology and Tectonic Settings
Eds. W. Nemec and R.J. Steel
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A spectrum of Middle Ordovician fan deltas and braidplain deltas, North Wales: a consequence of varying fluvial clastic input

G.J. ORTON

A spectrum of coarse-grained deltaic sequences has been recognized in the infill of a tectonically active, Middle Ordovician ensialic back-arc basin in North Wales. Fan deltas of three types (fluvial-dominated, wave-modified, and wave-dominated) are described, and are distinguished from a braidplain delta on the basis of depositional process, grain-size and facies gradients, and spatial relation to basin configuration. Coastal fans and braidplains differ in their effectiveness in coping with marine basin energy, in the resultant patterns of delta-front sedimentation, and in the type of associated coastline. The contrasting responses to the basinal marine regime, with development of either reflective or dissipative shorelines, can be related to variation in the supply rate and grain size of sediment delivered to the delta front. Variation in wave energy was of minor importance. Coarse-grained deltas can be classified according to processes dominating their fronts, but the ternary classification diagram should be extended in a fourth dimension to include sediment grain size along the axis of the dispersal system.

Introduction

In coarse-grained deltas, sediment transport is usually in high-gradient, high-discharge streams, often braided, deriving their load from small drainage basins. The grain size of sediment delivered to the delta front varies from medium sand to large boulders. Resultant deposits comprise subaerial, transitional, and subaqueous components.

The accessibility of delta-plain deposits on modern coarse-grained deltas has led to preferential documentation of the subaerial components of such systems. Several depositional settings have been recognized. These include alluvial-fan deltas, which range from stream- to mass flow-dominated (Sneh, 1979; Wescott and Ethridge, 1980; Schramm, 1981), coastal outwash plains (Boothroyd and Ashley, 1975; Hine and Boothroyd, 1978), bayhead deltas (Syvitski and Farrow, 1983; Prior *et al.*, 1986), and short-headed stream deltas (Flores, 1975).

A corresponding spectrum of delta types has not been recognized in ancient successions. The subaerial component, although an integral and important part of coarse-grained deltaic systems (Moore and Asquith, 1971; McGowen and Scott, 1974; Sykes and Brand, 1976), is not always well preserved, or simply proves difficult to interpret in terms of the geomorphic criteria derived from specific modern situations. Detailed studies have concentrated mainly on processes operating on the delta front (e.g. Pollard *et al.*, 1982; Farquharson, 1981; Van der Meulen, 1983; Nemec *et al.*, 1984; Massari, 1984; Postma, 1984; Surlyk, 1984). Comparative studies have related variation in the pattern of sediment dispersal and deposition in fan deltas to aspects of the receiving basin (e.g. Wescott and Ethridge, 1980; Ethridge and Wescott, 1984), rather than to the nature of particular alluvial system. Considerations of the subaerial component commonly failed to discriminate between fluvial systems of alluvial fans and those of braidplains, as pointed out by Rust (1979). As a result, braided streams building out into standing bodies of water were often treated as fan deltas, even where the deposit morphology and facies distribution suggest otherwise (e.g. Hobday, 1974; Galloway, 1976; Boothroyd and Nummedal, 1978; Harbaugh and Dickinson, 1981; Galloway and Hobday, 1983, p. 40; Kingsley, 1984; Pudsey, 1984). This overextension of the term 'fan delta', to encompass all coarse-grained deltas, has been noted by several authors (Sykes and Brand, 1976; Dixon, 1979; McPherson *et al.*, 1987, and this volume).

A fan delta *sensu stricto* represents progradation of an alluvial fan into a standing body of water (definition by Holmes, 1965; and McGowen, 1971). In ancient sequences, alluvial fans can be distinguished from braidplains on the basis of depositional process (the

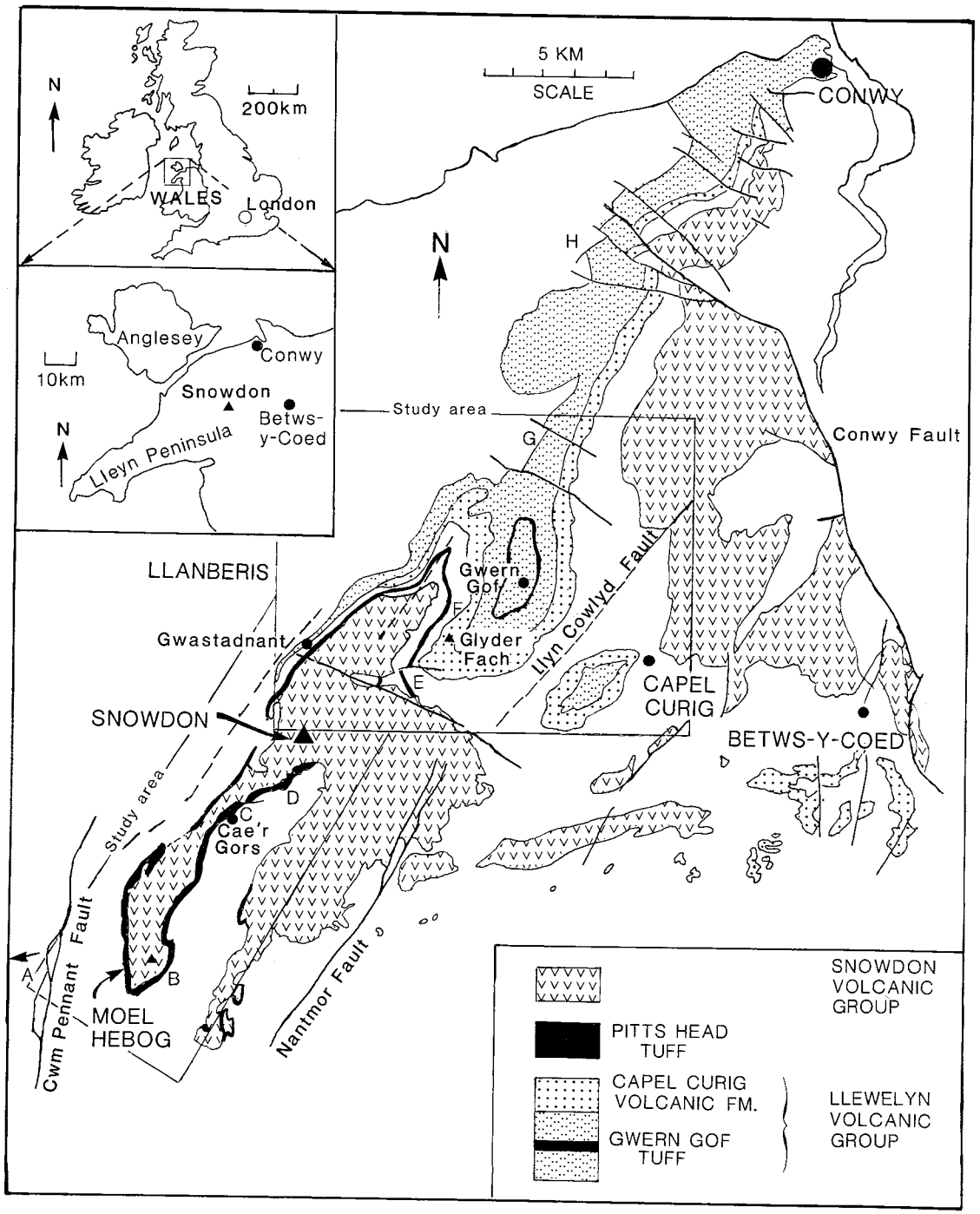

Figure 1. Location of study area and distribution of major Caradocian volcanic units, North Wales (after Reedman *et al.*, 1986). Letter symbols (*A–H*) indicate local sections shown in Fig. 2.

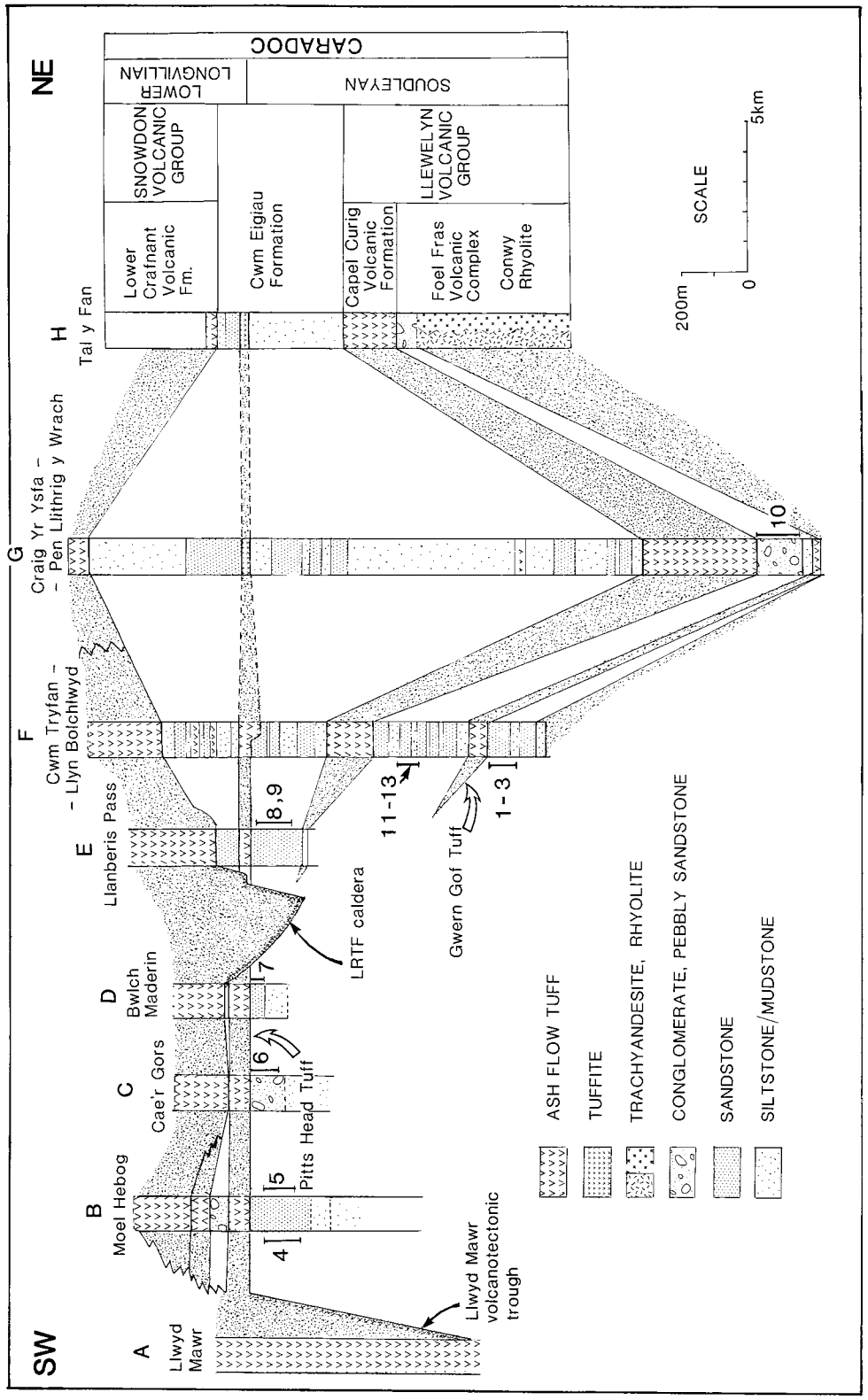

Figure 2. Generalized stratigraphical cross-section through the North Wales part of the Caradocian back-arc basin. Location of individual sections (*A–H*) is shown in Fig. 1. Their numbered segments indicate stratigraphical and regional position of the profiles (1–13) described in text, and shown in other figures.

likelihood of concentrated sheetfloods and debris-flows), radial palaeocurrents, rapid lateral and down-fan variation in grain size and attendant facies, and a transverse relationship to the structural grain, or configuration, of a basin (Bull, 1972; Rust, 1979). In contrast, braidplain deltas, as defined here, are associated with broad systems of low-sinuosity, braided streams which bear no strict affinity to alluvial fans. Modern examples include coastal outwash plains of Alaska or Iceland (e.g. Hine and Boothroyd, 1979). The term 'braid delta' has been proposed for such deposits (McPherson *et al.*, 1987, and this volume), but the term *braidplain delta* is preferred here to emphasize the deposit morphology as well as the nature of the contributing alluvial system. It is suggested that the term 'braid delta' be restricted to deltas whose braided distributary systems are merely derivative from the progradation of solitary, braided or low-sinuosity rivers; recent examples include bayhead deltas in the fjords in western Canada, short-headed stream deltas in Lake Erie (Flores and Ferm, 1970), and the Copper River delta (Galloway, 1976). The distinction between 'braid deltas' and 'braidplain deltas', as defined here, is important, as it has direct bearing on the differences in alluvial architecture, including the geometry of resultant sandstone bodies.

In this study, five coarse-grained deltas are reconstructed using a network of closely spaced vertical sections with laterally traced lithofacies in areas of good exposure. Emphasis is placed on the transition from alluvial plain to nearshore delta-front deposits, where interacting fluvial and marine processes resulted in subtle and complex facies differentiation. Fan deltas are distinguished from braidplain deltas by the geometry of the contributing alluvial system. Braidplains are basin axial systems fed laterally by alluvial fans, or showing no such association. It will be further demonstrated that the type and large-scale geometry of an alluvial system (fan *v.* braidplain), its internal behaviour (fixed *v.* mobile channels, channellized *v.* unchannellized peak-discharge) and sediment grain size, all have controlled the efficiency of discharge in coping with marine energy, the patterns of sediment dispersal under marine reworking, and the type of associated coastline.

Tectonic setting and stratigraphy

In Middle Ordovician time, the North Wales region was part of an ensialic marginal basin (Kokelaar *et al.*, 1984) of uncertain plate-tectonic setting. Volcanism within the basin was essentially bimodal, with voluminous eruptions of rhyolites and tholeiitic basalts. Crustal extension is indicated by magma composition and by the style of volcanicity and sedimentation. Rifting, however, was insufficient to produce an oceanic crust, and the basin developed on immature continental crust.

In North Wales, Ordovician sedimentation was dominated by muddy facies, with volcanicity confined to the Caradocian series. Two major volcanic successions have been recognized (Howells *et al.*, 1983): a lower, Llewelyn Volcanic Group of Soudleyan age and an upper, Snowdon Volcanic Group of Longvillian age (Figs. 1, 2, 3). Several hundred metres of tuffaceous sediments, deposited in a variety of environments, are intercalated within these volcanic successions. Laterally extensive rhyolitic ash-flow sheets, such as the Gwern Gof Tuff, Garth Tuff of the Capel Curig Volcanic Formation, and Pitts Head Tuff (Fig. 2), form useful markers for basin-wide (20 km) correlation of time-equivalent depositional sequences. Lines of structural weakness in the Precambrian basement were repeatedly activated as faults, controlling both the basin configuration and alluvial catchment areas. Related variation in the nature and geometry of fluvial drainage systems resulted in the observed spectrum of coarse-grained deltaic sequences. Four fan-delta examples are first described, and then are compared with each other and contrasted with a braidplain-delta example.

Fluvially dominated fan delta: the Gwern Gof depositional system

Sediments subjacent to the Gwern Gof Tuff (Fig. 5) comprise two large-scale (60 m) coarsening-upward cyclothems (*A*, *B*), each interpreted to represent an episode of delta progradation, separated and overlain by thin horizons of delta-abandonment facies. Only the lower (*A*) cycle (depositional phases 1–6 in Fig. 5) is described in detail.

The lowest, first coarsening-upward sequence

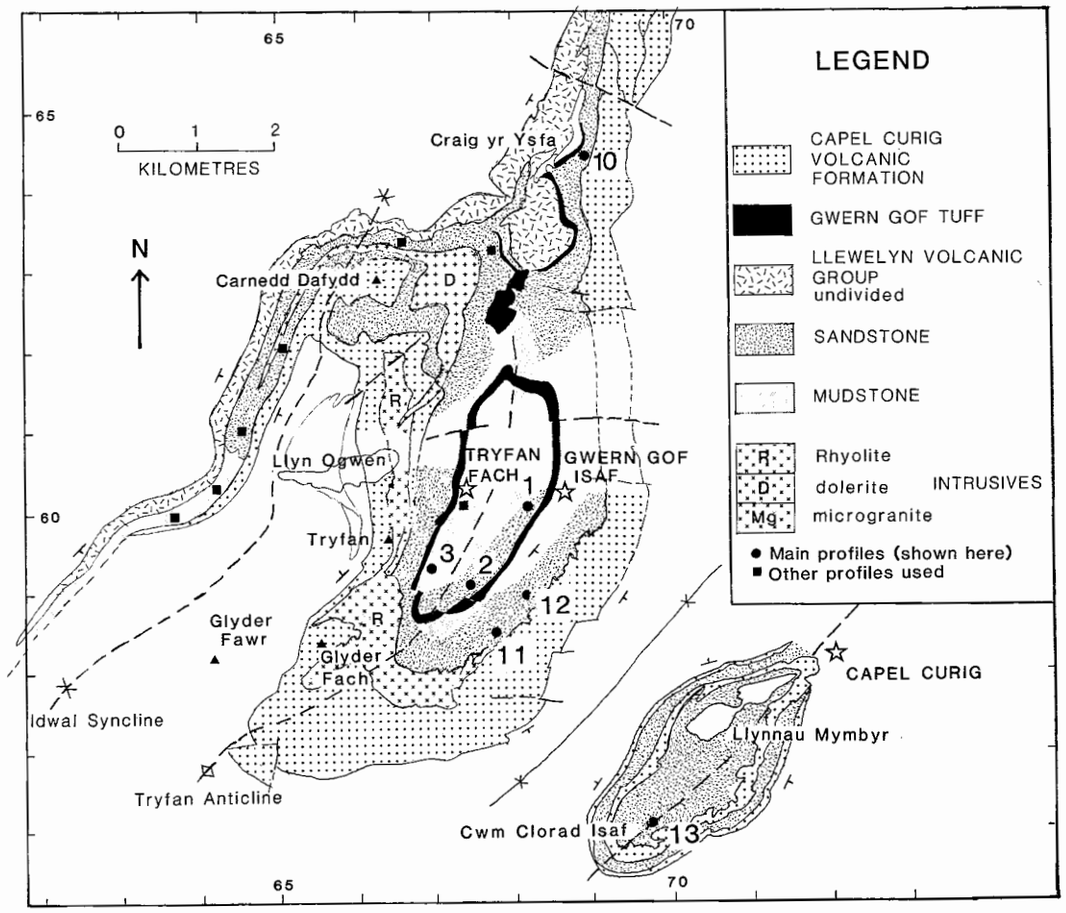

Figure 3. Geological map of the Tryfan anticline and surroundings, showing the distribution of Caradocian volcanics and associated clastics and the locations of main profiles (numbered dots) and other measured sections (black squares) used in this study. Numbers on map margins refer to national grid. Location of the area is shown in Fig. 1 (centre).

(Fig. 5, phase 2 in profile 2) is made up of thin (10–20 m), discrete packages of fine-grained sandstone which are thought to represent channel effluent-derived deposits (distributary mouth-bars), with inconstant sediment supply due to frequent channel avulsion. The inferred distal mouth-bar (Fig. 5, phase 2 in profile 3) is dominated by graded beds up to 20 cm thick. A subtle upward gradation from very fine to fine-grained sandstone is accompanied by thinning of beds to 10 cm in the proximal mouth-bar. These beds probably represent density currents from individual river-flood events. The second coarsening-upward mouth-bar sequence (Fig. 5, phase 7) indicates a larger, more rapid supply of sediment into the nearshore zone. Distal bar deposits are still pervasively bioturbated, but the more proximal sandstones display distinct horizontal bedding, with sand layers 5–8 cm thick (Fig. 7*A*). Occasional wave-ripple lamination and thin (10 cm) sets of sigmoidal cross-strata (?tidal features) indicate minor marine influence.

A fining-upward unit of cross-stratified medium-grained sandstone (Fig. 5, profile 2, phase 4), overlying the first delta-front sequence, represents infill of a broad, low-sinuosity distributary channel. Scour relief indicates that the channel cut at least 4.5 m into a thin (less than 8 m), laterally continuous (extent more than 500 m) unit of fine-grained deposits which coarsen upward from mudstone to coarse siltstone (Fig. 6).

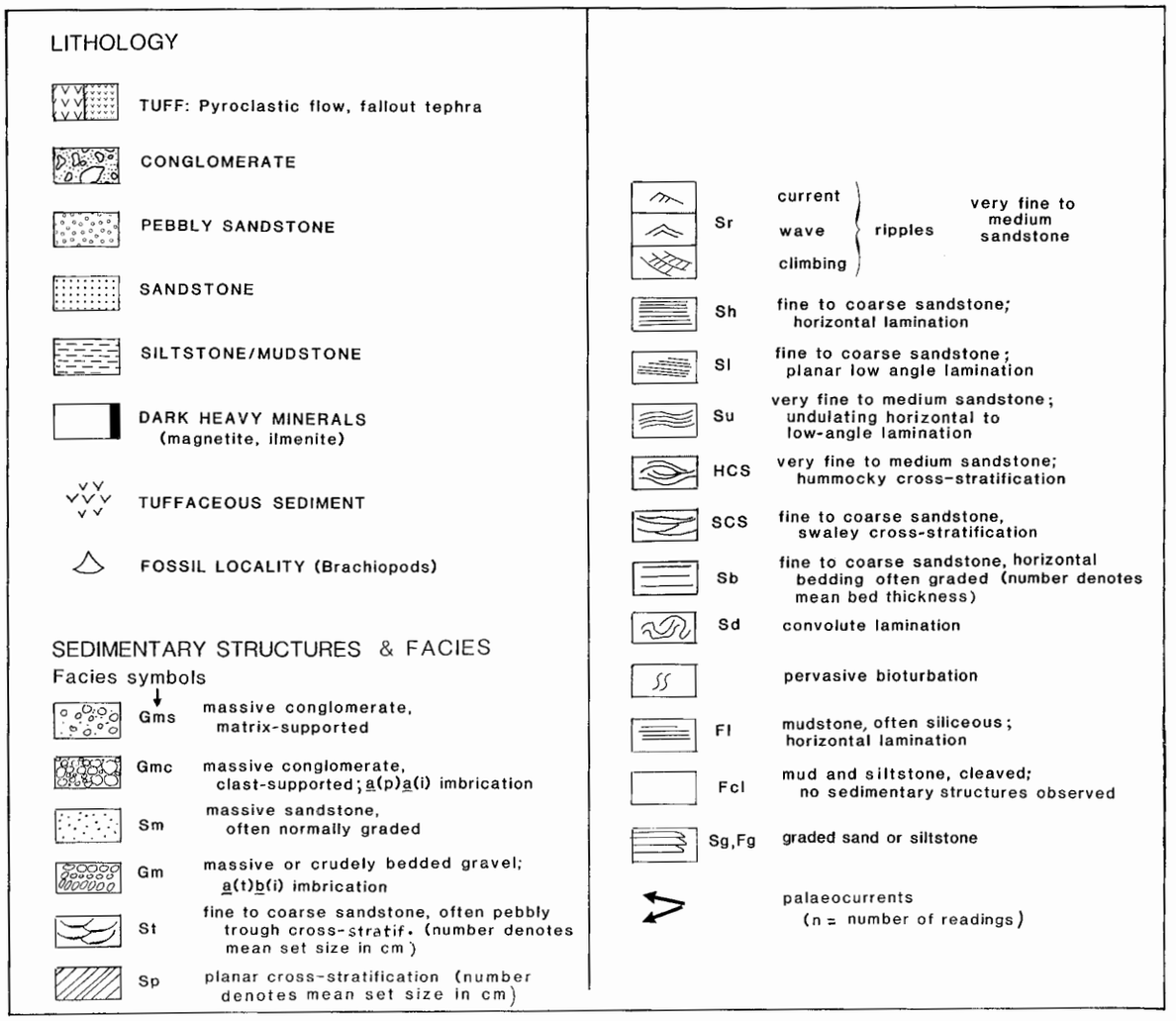

Figure 4. Legend to the stratigraphical profiles shown in the other, following figures. Facies letter-code is modified from Miall (1977).

Figure 5. Stratigraphical profiles through the Gwern Gof fan-delta succession (see profile locations in Fig. 3 and legend in Fig. 4). Time equivalence is implied for the numbered phases of deposition (large numbers (*1*) to (*8*) on profile margins). Inset map shows interpreted palaeogeography during maximum progradation (phase 4) of the lower (*A*) delta; black squares denote other profiles used for this reconstruction. Palaeogeography is not corrected for NW–SE tectonic shortening of 40%.

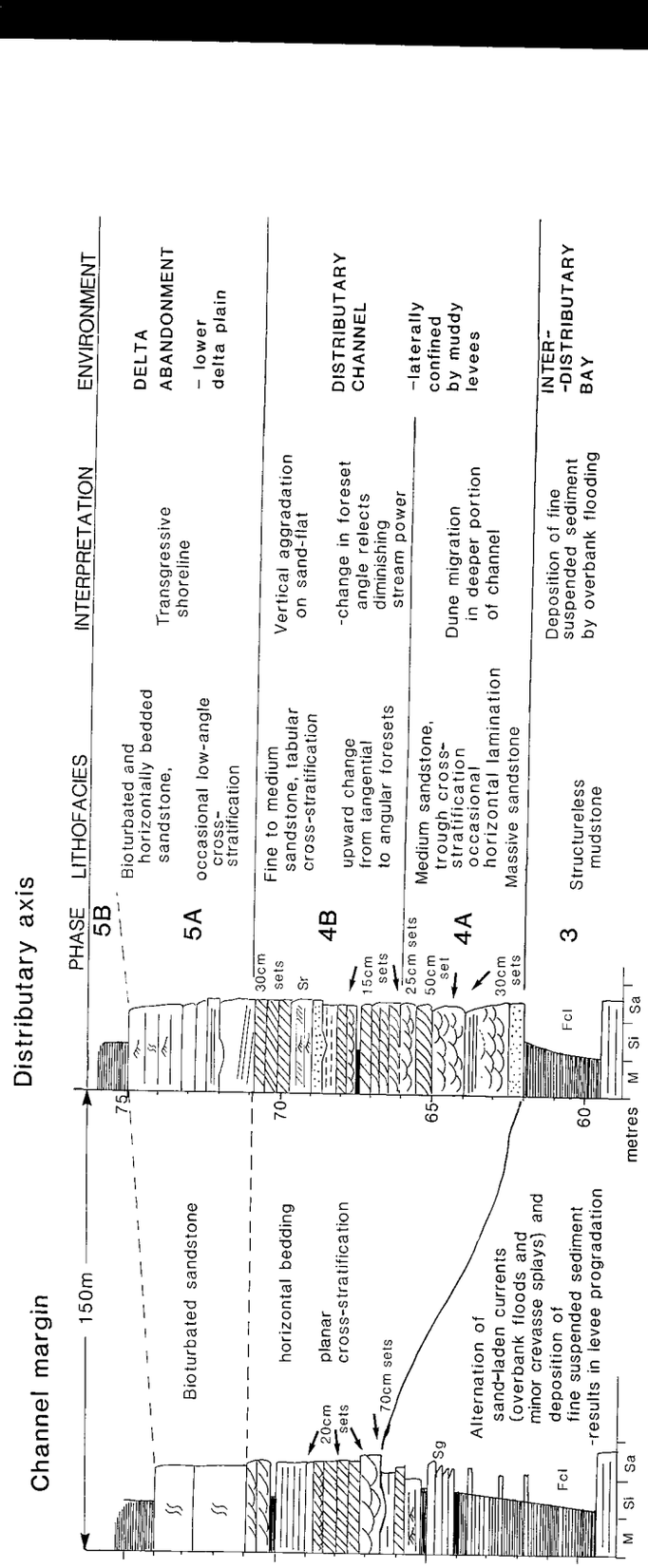

Figure 6. Details of a distributary channel-fill sequence from the Gwern Gof fan delta (see Fig. 5, phase 4 in profile 2).

This subjacent unit probably represents a broad interdistributary bay, with occasional thin (15 cm) interbeds of graded and/or horizontally laminated sand reflecting overbank flooding (Elliott, 1974). The channel itself was at least 9 m deep (infill thickness) and 150 m wide (outcrop extent). Its infill fines upwards from comparatively well-sorted, medium-grained to fine-grained sandstones. Tabular planar cross-stratification (phase 4*B* in Fig. 6), with tangential to angular foreset strata and unidirectional dip azimuths towards the west, is the dominant sedimentary structure (Facies *Sp*). Vertically stacked cross-sets suggest that the channel was filled by rapid aggradation of coalescent transverse bars, or 'sand flats' (e.g. Smith, 1972), rather than by lateral accretion.

More proximal fluvial deposits (Fig. 5, phase 4 profile 1) include trough and planar cross-stratified (Facies *St* and *Sp*) medium/coarse pebbly sandstones. These facies and ripple cross-laminated sandstones (Facies *Sr*) are organized into fining-upward sequences, 1–2 m thick, suggesting deposition in much shallower braided streams.

The cessation of sediment supply to the first delta lobe was associated with an episode of distal volcanicity in the source area. The delta plain is capped by fine-grained facies (Fig. 5, profile 1, top of phase 4) containing units of airfall tuff interbedded with thin debris-flow units. These deposits, the radial pattern of fluvial palaeocurrents (cf. profiles 1 and 2 in Fig. 5), and a transverse relationship to the basin margin suggest that the fluvial system was associated with a large alluvial fan, probably reflecting uplift on the Conwy fault, 10 km to the E (Fig. 1). The abundant supply of mud, derived by chemical weathering of subjacent volcanics, forced the distributary braided streams to form deeper channels on the lower delta-plain. Galloway (1981) has described similar, bedload streams of the Cenozoic Guedyan fluvial system which were confined due to deposition of cohesive, rapidly altered volcanic ash.

Delta-abandonment facies vary according to position on the delta front. Bioturbated, fine-grained sandstones (Fig. 6, phase 5*A*) represent marine-reworked remnants of distributary channel deposits. The overlying mudstone (phase 5*B*) extends laterally for 1500 m. It contains rare thin sandstone interbeds and is interpreted as a back-barrier lagoon deposit. An overlying, 5 m thick sequence of well-sorted, magnetite-rich, fine- to medium-grained sandstones (Fig. 5, phase 6) with medium-scale (20 cm) trough cross-stratification and horizontal lamination is interpreted as a barrier island deposit. Beachface facies and heavy-mineral concentration indicate considerable nearshore wave activity in the absence of fluvial input. Similar facies have been described from modern, non-barred, high-energy coastlines of California and Oregon (Clifton *et al.*, 1971; Howard and Reineck, 1981).

Interpreted palaeogeography during maximum progradation of the fan delta (Fig. 5, inset map) indicates a lobate geometry of the delta front and a moderate extent of basinward sediment transport. The continuous delta-front sand prism resulted from lateral coalescence of the lobate 'unit' mouth-bars of closely spaced distributary channels. As the delta facies are dominantly fluvial or effluent-derived, despite the considerably high wave-energy levels inferred, the Gwern Gof fan delta is categorized as a fluvially dominated variety.

Wave-modified fluvial fan delta: the Moel Hebog depositional system

At Moel Hebog (Fig. 2, column *B*), coarsening-upward sequences (Fig. 8) represent stacking of successively more proximal deltaic facies in response to both faulting and fan-lobe progradation. Axial mouth-bar sequences (Fig. 8, phase 2 in profile 4) coarsen upwards to medium-grained sandstones and abruptly overlie mud- and silt-dominated prodelta deposits (phase 1). Mouth-bar sandstones comprise a variety of facies (Fig. 8), including amalgamated sets of hummocky cross-stratification (Fig. 7*B*) which indicate strong nearshore activity of storm waves. Several beds (up to 1.5 m thick) display contorted lamination and large fluid-escape structures, probably due to gravity-driven deformation of the delta slope.

Delta-plain deposits (Fig. 8, phase 4 in profile 4) are dominated by trough cross-stratified, medium to very coarse pebbly sandstones which are interpreted to have been deposited in relatively small, unstable braided streams. Channel-fills are crudely recognizable as 3–5 m thick, fining-upward sequences. Planar cross-stratification is rare, probably owing to constant shifting of channels (pools) and the accompanying destruction of channel bars (Campbell, 1976). The common occurrence of debris-flow beds in the overlying, more proximal alluvial deposits (e.g. Fig. 8, phase 5 in profile 5) suggests that the fluvial system

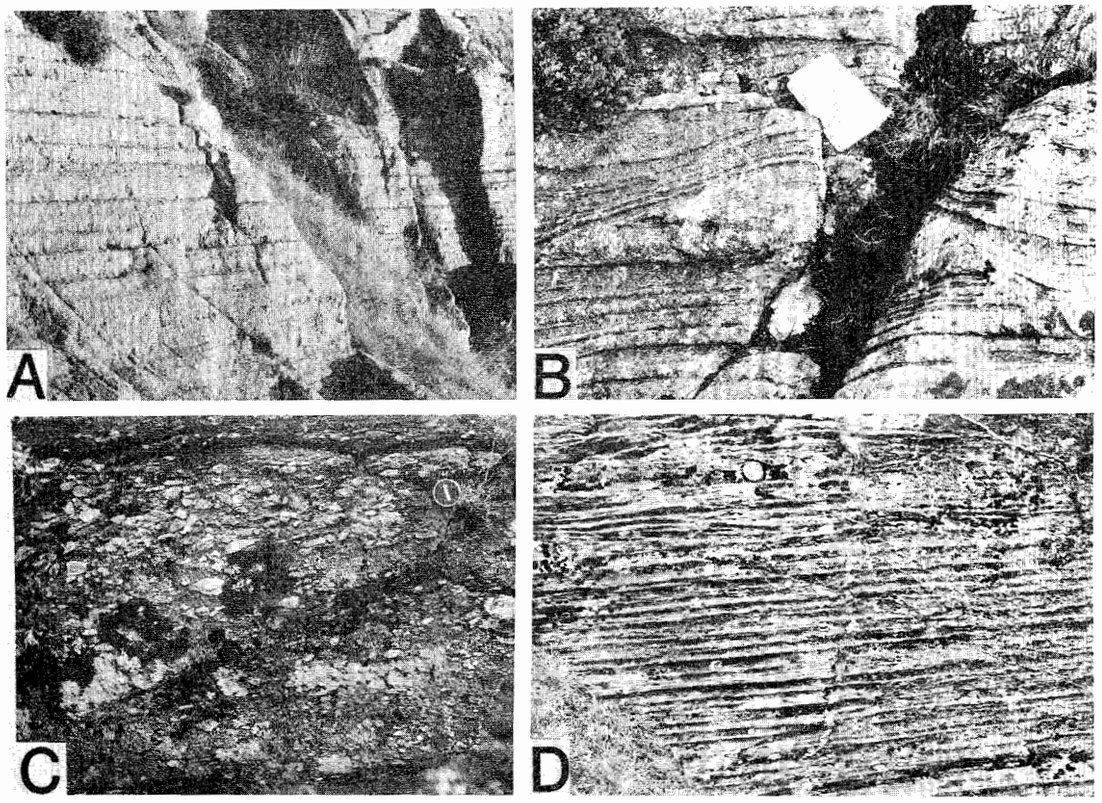

Figure 7. Facies details from the fan-delta successions. (*A*) Normally graded beds of medium/fine sandstones, 5–15 cm thick, interpreted as stream-effluent flood deposits of a proximal mouth-bar (from phase 7 in profile 2, Fig. 5). (*B*) Hummocky cross-stratification typical of wave-reworked mouth-bar sands (from phase 2 in profile 4, Fig. 8); notebook is 18 cm long. (*C*) Inversely graded, matrix-supported bed of debris-flow conglomerate from the delta-plain deposits of Cae'r Gors fan delta (phase 4 in profile 6, Fig. 9); lens cap is 6 cm. (*D*) Horizontal and low-angle lamination in magnetite-rich, fine-grained swash-bar sandstone deposited some 3 km (longshore) off the axis of Cae'r Gors fan delta (from phase 3 in profile 7, Fig. 9); lens cap is 6 cm.

operated in the middle to distal segment of an alluvial fan.

The fan-delta fringe zone was the most complex depositional environment. The very fine- to medium-grained sandstones of phase 3 (Fig. 8) are thought to represent a moderate wave-energy shoreface which formed part of a barrier spit. Sedimentary structures include plane parallel lamination with low-angle truncations. Occasional sets of trough and planar cross-strata indicate palaeocurrents towards the NE (inferred offshore).

An anomalous, 25 m thick heterolithic interval (Fig. 8, phase 5 in profile 4) is interpreted as representing a tidally influenced lagoon between coastal fan salients (lobes) which was protected from strong wave action by a barrier spit or swash bars (cf. Vos and Eriksson, 1977; Vos, 1981). Mudstone and siltstone are interbedded with coarse sandstones (0.5–50 cm units) and contain thin (2–5 cm) layers of airfall tuffs. Heterolithic flaser/lenticular bedding involves wave ripples and indicates bimodal palaeocurrents. Sandstone interbeds display plane parallel lamination, low-angle cross-stratification and climbing-ripple cross-lamination, whereas thicker beds (up to 2 m) contain medium-scale trough cross-stratification. These units represent crevasse splays or storm washover sheets and creavasse channels, respectively. The facies assemblage is somewhat similar to the lacustrine/floodbasin and transitional fan-fringe sequences described from the Hornelen Basin (Steel and Aasheim, 1978).

In adjacent exposures (e.g. profile 5 in Fig. 8), the

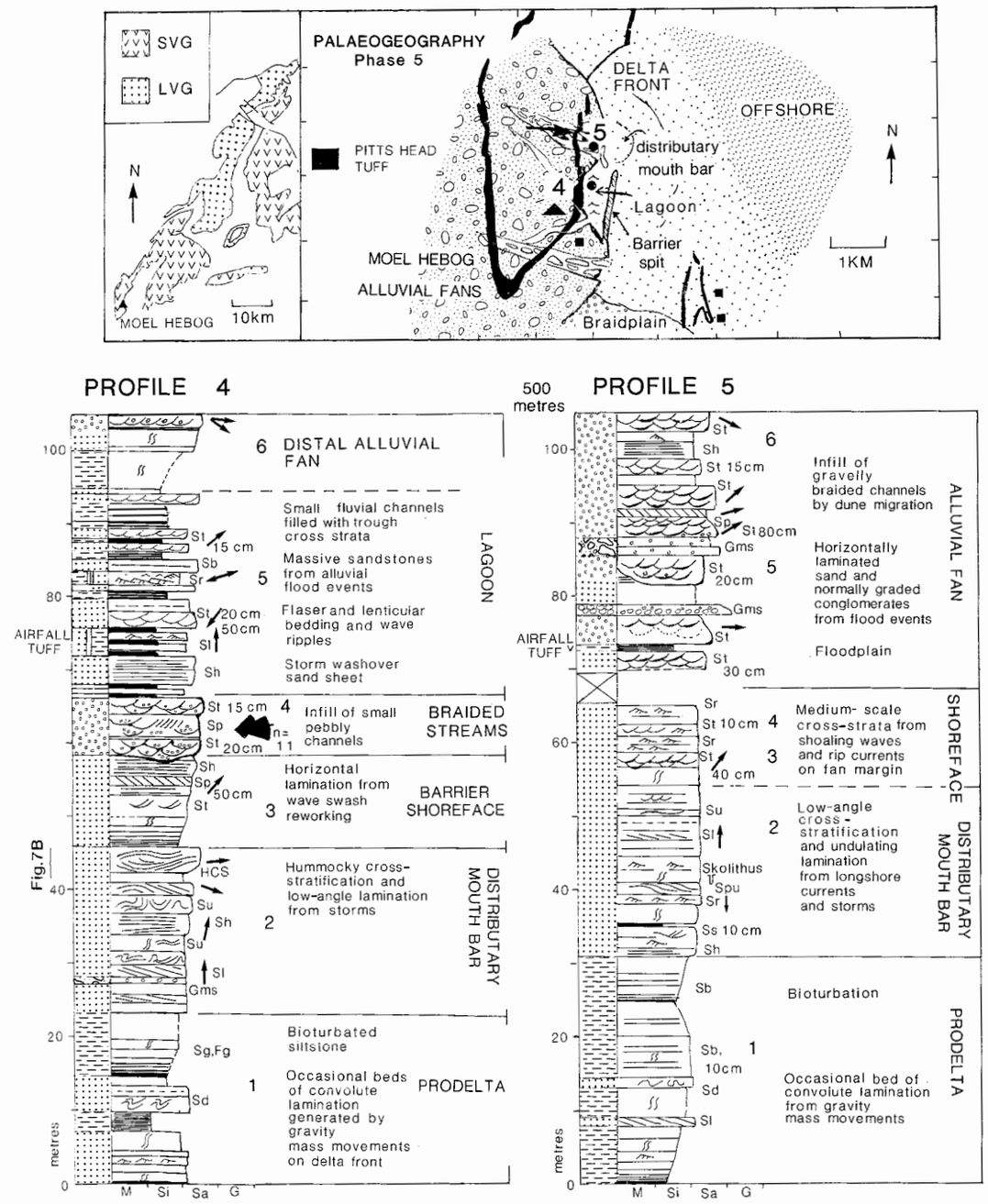

Figure 8. Profiles through the Moel Hebog fan-delta succession (see legend in Fig. 4). Inset map shows location of profiles and interpreted palaeogeography (phase 5 of fan-delta evolution); black squares indicate location of other profiles used in this reconstruction.

airfall tuffs occur as angular rip-up clasts within pebbly, trough cross-stratified sandstones deposited in poorly channelized braided streams. It appears that tectonic changes, associated with the volcanicity, caused sideways migration of the alluvial system and the locus of sediment input into the basin. Interpreted palaeogeography at this time period (phase 5) indicates coalescing coastal fan-lobes at least 2 or 3 km wide (Fig. 8, inset map). With the fan-lobe radius assumed to have similar dimension, the alluvial system reflects uplift on the Cwm Pennant or related faults to the west (Fig. 1). Facies sequences indicate coastal fan progradation into a basin with high nearshore wave energy, although the sediment dispersal patterns suggest rather moderate reworking by waves. Accordingly, the Moel Hebog palaeosystem is categorized as a fluvially constructed, wave-modified fan delta.

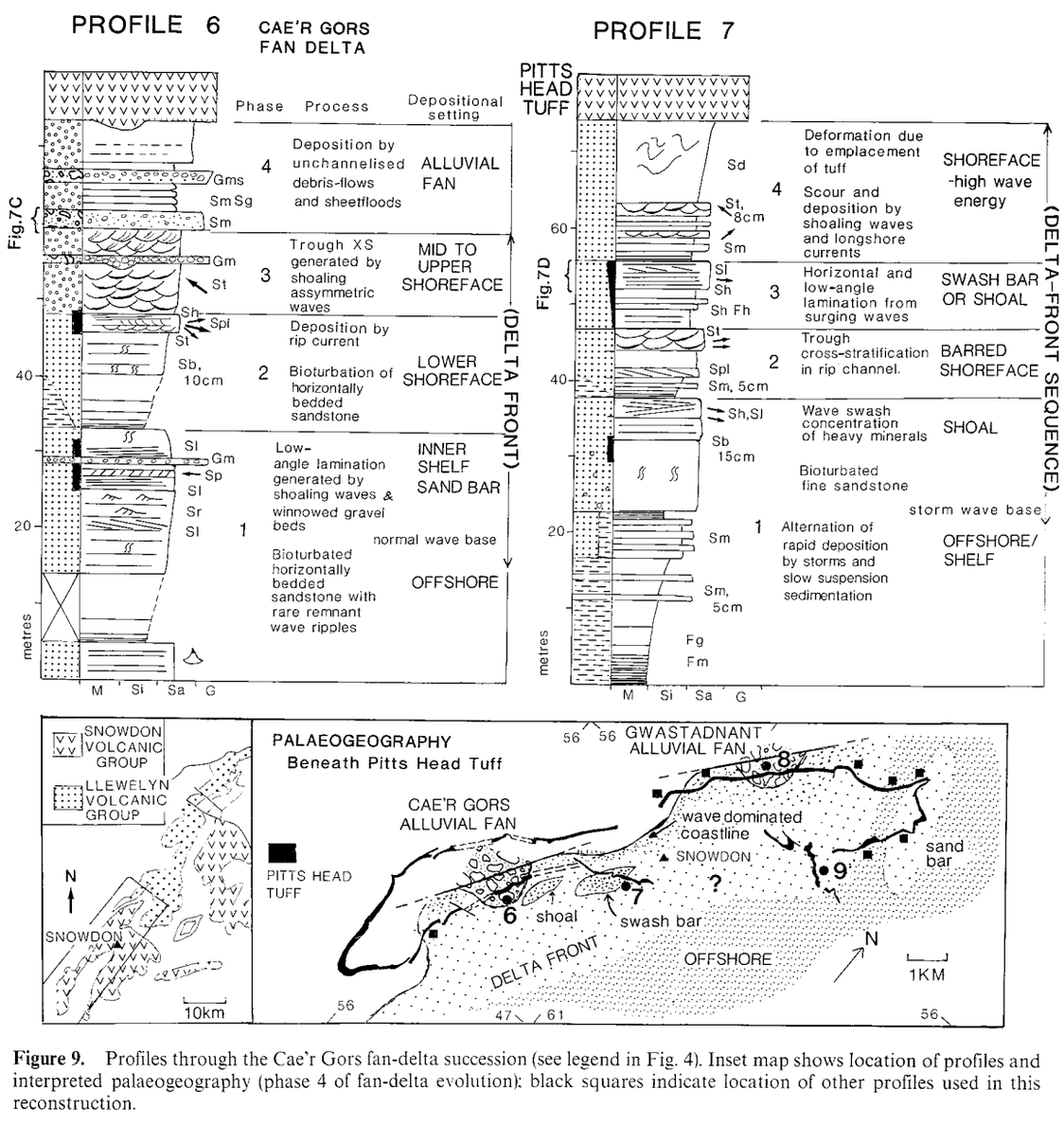

Figure 9. Profiles through the Cae'r Gors fan-delta succession (see legend in Fig. 4). Inset map shows location of profiles and interpreted palaeogeography (phase 4 of fan-delta evolution): black squares indicate location of other profiles used in this reconstruction.

Wave-dominated fan deltas: the Cae'r Gors and Gwastadnant depositional systems

Deposits of inferred wave-dominated fan deltas (Figs. 9, 10) are characterized by thin alluvial sequences (10 m) of conglomerates and pebbly sandstones overlying shoreface-type sequences (30–70 m thick) of fine- to medium-grained sandstones. Four successions from two independent depositional systems, at Cae'r Gors and at Gwastadnant (Fig. 1), are described for illustration. The deposits occur directly below the Lower Pitts Head Tuff ash flow sheet (Figs. 9, 10).

At Cae'r Gors (Fig. 9, profile 6), the succesion comprises two coarsening-upward sequences, both representing progradation of a high wave-energy shoreline dominated by shoaling-wave processes (cf. Clifton *et al.*, 1971; Hunter *et al.*, 1979; Clifton, 1981). The first sequence (phase 1) coarsens up to medium sandstone, dominated by 1–2 cm thick sets of horizontal or low-angle lamination with ubiquitous magnetic heavy-minerals. Absence of trough cross-stratification may be a function of finer grain sizes. The observed stratification types are attributed to high-power (relative to grain size) oscillatory currents, and are thought to represent bedforms superimposed on low-amplitude, swash-formed bars. Similar deposits have been described from the margins of both modern and ancient fan deltas (McGowen, 1971: Vos and Eriksson, 1977). The second delta-front sequence (Fig. 9, phase 2 in profile 6) represents new, more pronounced progradation of the system. It coarsens upwards to pebbly, medium/coarse sandstones with some associated conglomerates (profile 6). Sandstones display plane-parallel lamination, trough cross-stratification and planar tabular cross-stratification with tangential foresets. Offshore (SE) palaeocurrents are dominant and probably represent redistribution of delta-front sediments by rip currents.

At Gwastadnant (Fig. 10, profile 8), the composite delta-front sequence represents similar, repetitive progradation of a shoreface (delta-front) environment. The deposits are bioturbated siltstones and very fine-grained sandstones. Sedimentary structures are largely obscured; remnants of thin (10–20 cm) bedding, rare horizontal lamination and ripples are the only recognizable features. Periodical influx of coarser sediment is indicated by thin graded beds of very coarse- to medium-grained sandstone. The sequence is thought to represent distal to medial mouth-bar deposits, heavily affected by bioturbation. They are unconformably overlain by thin, strongly erosive alluvial sequence (Fig. 10, phase 2*B* in profile 8).

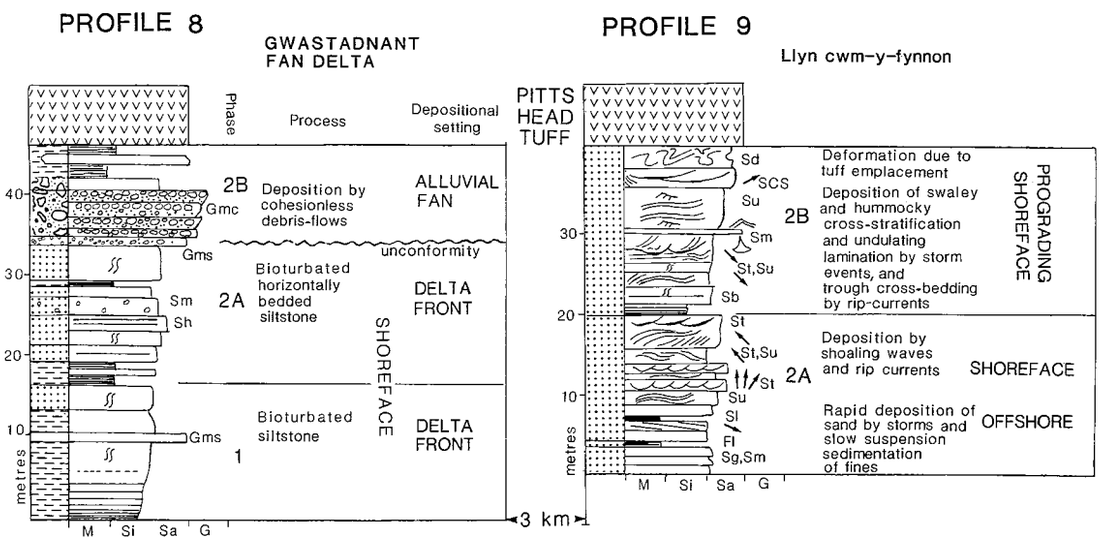

Figure 10. Profiles through the Gwastadnant fan-delta succession (see legend in Fig. 4 and profile locations in Fig. 9). Interpreted palaeogeography (phase 2*B*) is shown in the inset map in Figure 9.

The alluvial deposits in both fan deltas (Figs. 9, 10) indicate deposition by sediment gravity flows in largely unchannellized distributary systems. However, the conglomeratic facies vary between the two alluvial fans:

(i) At Cae'r Gors (Fig. 9), matrix-supported conglomerates dominate (Fig. 7*C*). These are either ungraded or inversely (coarse-tail) graded, with gravel fraction comprising less than 10% of bed volumes. Clasts are moderately well rounded and have a maximum size of 10 cm. Maximum bed thickness is 100 m.
(ii) In contrast, clast-supported conglomerates dominate the alluvial fan at Gwastadnant (Fig. 10). Beds are inversely graded to ungraded, often with matrix-rich bases, and occasionally display *a*(p)*a*(i) clast fabric. Maximum observed clast size is 33 cm. Bed thicknesses range from 80 to 200 cm.

Both conglomerate facies are interpreted as debris-flow deposits. The beds are attributed to laminar sediment flowage with effective shearing restricted to the flow base (ungraded deposits) or a thicker lower zone (inverse grading); larger clasts probably settled out from the shearing zone due to the loss or reduction of strength in the flow matrix. The beds show no evidence of channelling on an outcrop scale (20 m). The textural difference between mass-flows in the two alluvial fans is attributed to their derivation from different source areas (Fig. 9, inset). The debris-flow beds are often associated with unstratified, normally graded units (10–15 cm thick) of medium to coarse, pebbly sandstones. These are interpreted to have been deposited subaerially by fluidal sediment flows, or hyperconcentrated 'sheetfloods', where turbulence and high sediment concentration (high apparent viscosity) were probably the dominant grain-support mechanism (cf. Nemec and Steel, 1984).

In the two fan deltas, the demarcation between subaerial and subaqueous facies is not obvious. A thick sequence of trough cross-stratified, medium-grained sandstones underlying the alluvial deposits in the Cae'r Gors sequence (phase 3 in Fig. 9) could represent either braided stream or upper shoreface environments. The evidence of shoreward (NW) palaeocurrents favours the second interpretation. Assuming that alluvial deposits are solely represented by debris-flow and sheetflood facies, the mapped lateral extents of these facies (*c.* 800 and 300 m in the Cae'r Gors and Gwastadnant areas, respectively) indicate that the two fans were very narrow (Fig. 9, inset map).

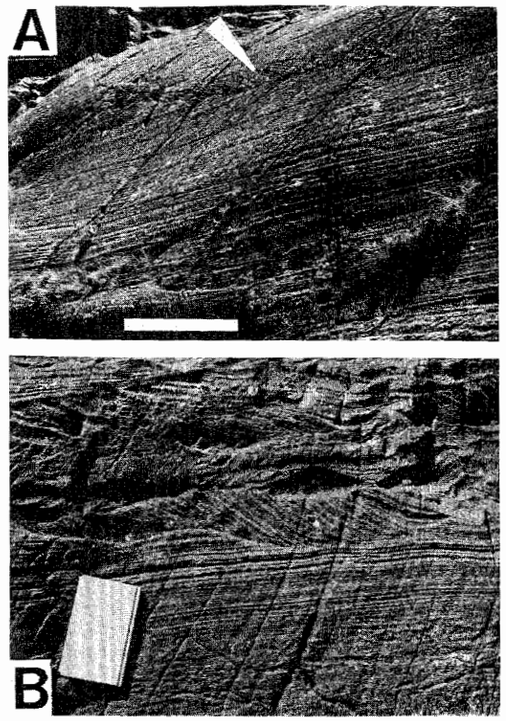

Figure 11. Facies details from the delta-front (shoreface) sequence in the Llyn Cwm-y-fynnon area (phase 2*A* in profile 9, Fig. 10; for location see inset in Fig. 9). (*A*) Low-angle, seaward-dipping planar to undulatory lamination in medium-grained sandstone. Note the intervening solitary set of medium-scale trough cross-strata (arrow). Scale bar is 50 cm. (*B*) Sets of medium-scale trough cross-strata attributed to shoreface reworking by asymmetrical shoaling waves (same sequence as above). Notebook is 18 cm long.

The patterns of sediment dispersal on the delta front are revealed by analysis of stratigraphically equivalent strata in adjacent profiles. Approximately 3 km off the Cae'r Gors fan axis, along the depositional strike, the Pitts Head Tuff is underlain by a thick (100 m), fine-grained, coarsening-upward succession (Fig. 9, profile 7). Its upper, sandy portion is organized into four coarsening-upward sequences, each reflecting an episode of shoreline progradation. Sandstones in the upper parts of these sequences display parallel, seaward-dipping (ESE) lamination and low-angle planar and shallow trough cross-strata, all containing abundant heavy minerals (Fig. 7*D*). These facies indicate a nearshore environment with high wave-energy and considerable periods of sediment reworking by wave swash (cf. Reid and Frostick, 1985). This regres-

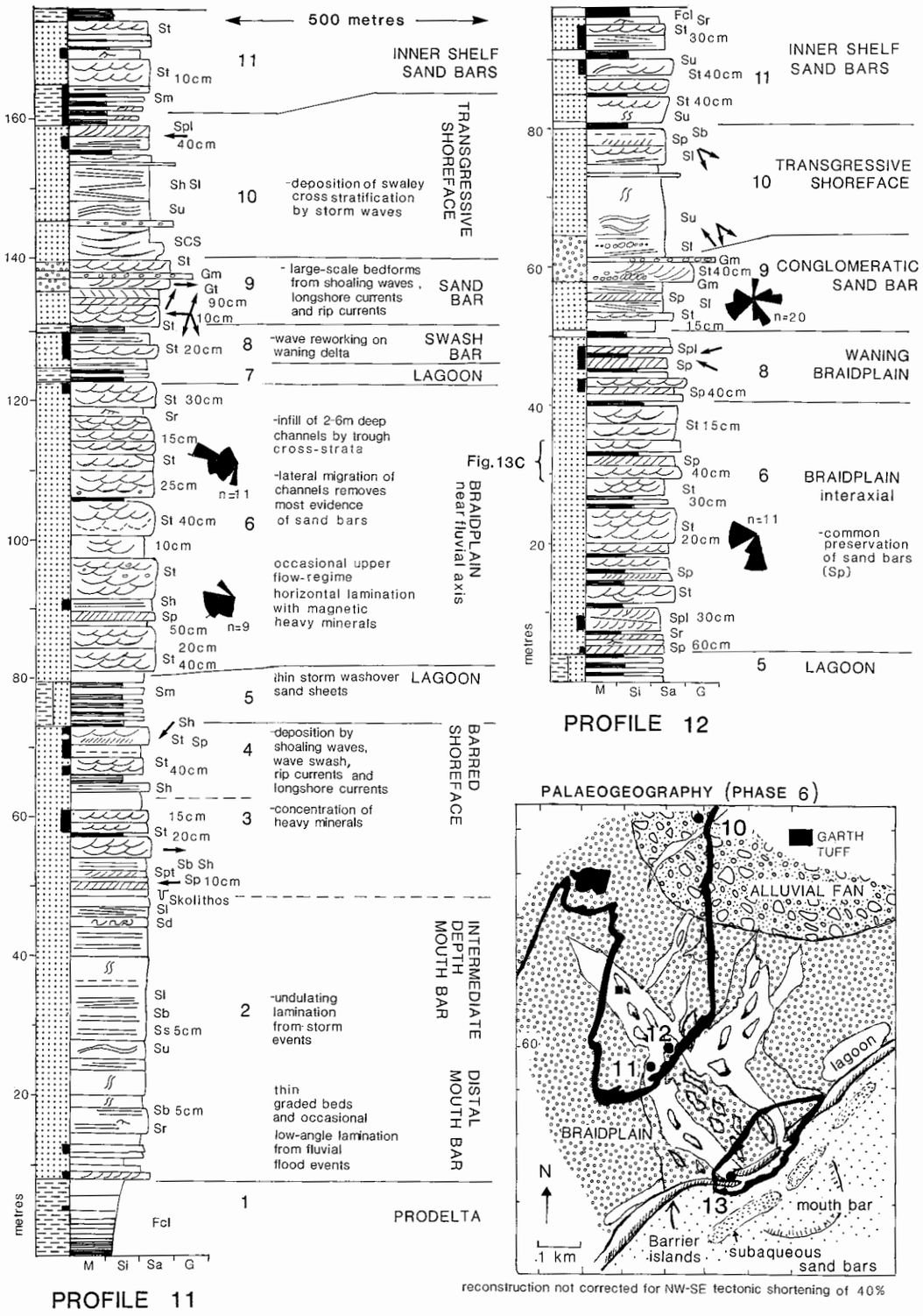

Figure 12. Profiles through the Capel Curig braidplain-delta succession (see location in Fig. 3 and legend in Fig. 4). Inset map shows interpreted palaeogeography (phase 6 of delta evolution); profiles 10 and 13 are shown in Figs. 14, 15.

sive succession could either represent wave-driven longshore drift of material derived from the Cae'r Gors fan delta, or be a distal segment of another wave-dominated delta lobe (cf. Heward, 1981). Correlative cyclicity (Fig. 9), similar heavy-mineral contents and the short distance involved favour this former interpretation.

At Llyn Cwm-y-fynnon (Fig. 10, profile 9), some 3 km basinwards off the Gwastadnant fan, the Pitts Head Tuff is directly underlain by a coarsening-upward sandy succession (phase 2) interpreted to represent progradation of a storm wave-dominated shoreline. Sediments of the lower/middle shoreface, immediately subjacent to the Pitts Head tuff, comprise alternations of undulatory horizontal lamination and low-angle planar or swaley cross-stratification in medium-grained sandstones, but display shallow trough cross-strata or straight-crested wave megaripples in very coarse sandstones (Fig. 11; see also similar features in Fig. 13*B*, *D*). This differential response to high-energy shoaling waves was probably a function of sediment grain size. Available palaeocurrent data suggest that the shoreline prograded due to redistribution of sediments delivered to the Gwastadnant delta front. Both this depositional system and the one at Cae'r Gors (see above) thus represent wave-dominated fan deltas.

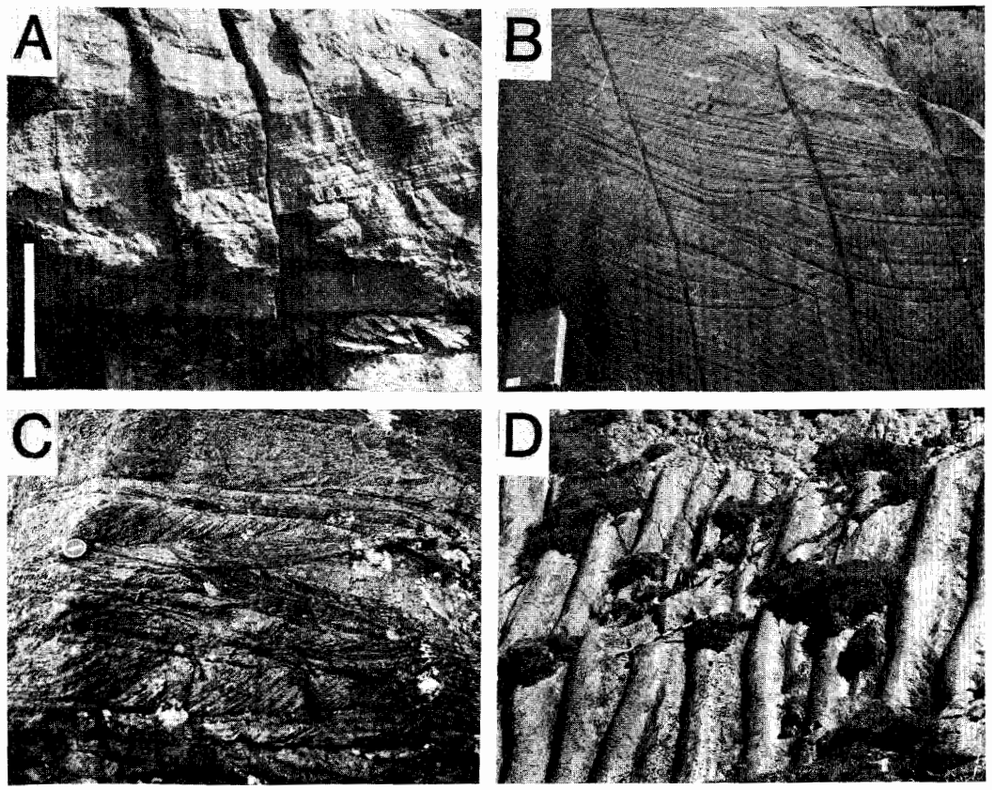

Figure 13. Facies details from the Capel Curig braidplain-delta succession. (*A*) Solitary set of low-angle, tangential cross-strata, attributed to small distributary channel, cut into the underlying, wave-reworked deposits of a mouth-bar (detail from profile 13, Fig. 15). Scale bar is 1 m. (*B*) Swaley cross-stratification in medium-grained sandstone (detail from same profile). Notebook is 18 cm long. (*C*) Coset of unidirectional trough cross-strata from the infill of braided fluvial channel (detail from phase 6 in profile 12, Fig. 12). Lens cap is 6 cm. (*D*) Large wave-ripples in very coarse sandstone (from profile 13, Fig. 15). Oblique view of a bedding surface; measuring stick (centre) is scaled in 10-cm bars.

Wave-modified braidplain delta: the Capel Curig depositional system

Sediments below the Capel Curig Volcanic Formation (Fig. 3) represent alluvial braidplain systems which prograded towards the S and SE and were fed laterally by alluvial fans related to an uplift along the Conwy fault system (Fig. 1). Stratigraphical variations in sediment composition (including rhyolitic detritus and magnetic heavy minerals), occurrence of airfall tuffs and the relative stratigraphical position of sedimentary successions allow tentative correlation of the alluvial-fan sequences with those of adjacent braidplain systems. Only one braidplain delta, at Capel Curig, and its contributing alluvial fan at Craig-yr-Ysfa, are described below for illustration.

The Capel Curig braidplain-delta succession (Fig. 12) represents progradation, strong aggradation, and eventual retreat of a fairly persistent depositional system. The delta-front sequence (phases 1–2) coarsens upward from mudstones to medium-grained sandstones and is thought to represent progradation of a distributary mouth-bar complex. Distal (lower) facies are very fine-grained sandstones showing thin (10–15 cm) graded beds, ungraded horizontal layering (1–5 cm), low-angle planar cross-strata, and occasional ripple cross-sets. More proximal (overlying) facies are medium-grained sandstones with undulatory to horizontal stratification (3–10 mm laminae), reflecting higher levels of both effluent energy and marine reworking.

This sequence (70 m) is overlain by two thinner, more complex sequences (phases 3–4 in Fig. 12) attributed to progradation of a barred shoreface. The sediments are siltstones to medium-grained sandstones in alternating beds. Sandstones display medium-scale trough cross-stratification (sets around 40 cm thick) and planar horizontal lamination, both with concentrated heavy minerals, indicative of high nearshore wave-energy. The facies show no consistent vertical arrangement (Fig. 12), probably reflecting numerous episodes of shoreline progradation and retreat.

These deposits are overlain by a thin (4 m), laterally persistent (more than 500 m) unit of interbedded very fine-grained, horizontally laminated sandstones (10–20 cm beds) and siltstones (phase 5 in Fig. 12). They are interpreted as storm washover sheets in a muddy back-barrier lagoon. The subjacent shoreface sequences would then be part of a barrier-island complex.

The overlying alluvial sequence (phase 6 in Fig. 12) consists of moderately well sorted, mainly medium-grained sandstones in broad tabular units (4–7 m

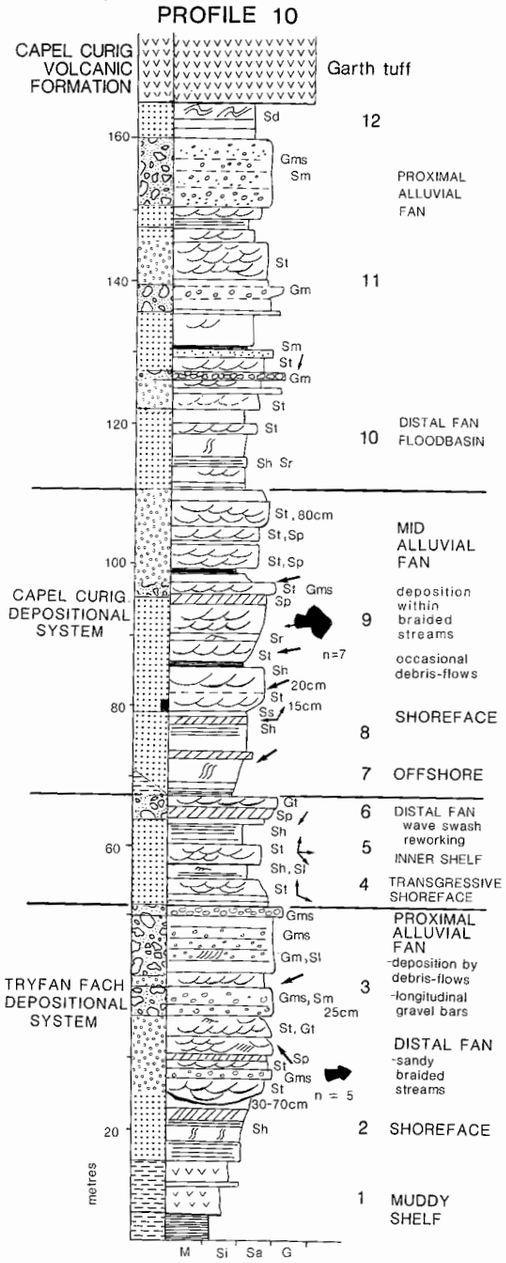

Figure 14. Profile through the basin-margin alluvial-fan succession at Creig-yr-Ysfa (see profile location in Fig. 12 and legend in Fig. 4). Phases 1–3 can be correlated with the development of Tryfan Fach braidplain delta (not discussed), whereas phases 7–9 indicate a link with the Capel Curig braidplain (see maps in Figs. 3, 12).

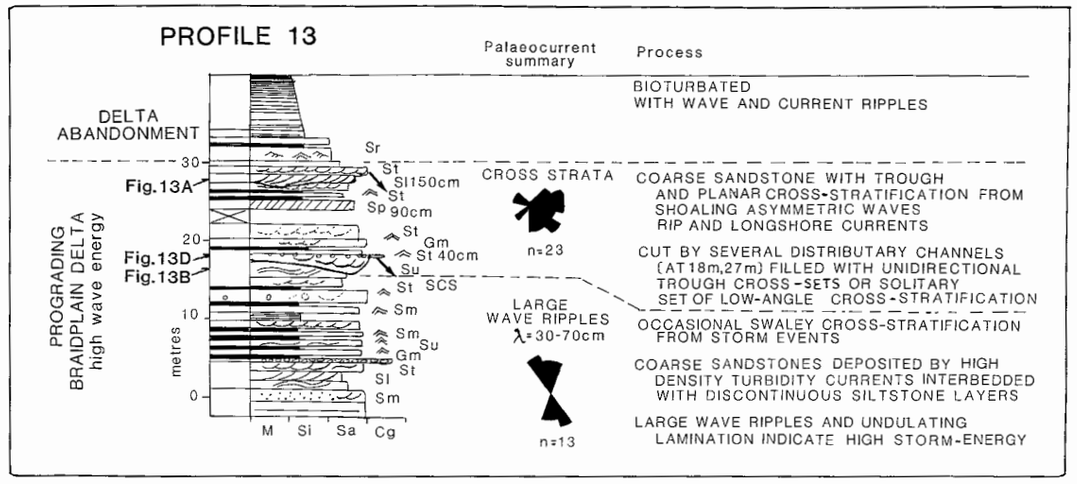

Figure 15. Profile through the distal segment of Capel Curig braidplain delta (see profile location on Fig. 12 and legend in Fig. 4).

thick) separated by thin siltstones. Units have sharp, erosive bases and are often capped by a thin horizon of horizontally laminated or ripple cross-laminated, very fine-grained sandstone. Trough cross-stratification dominates (Fig. 13*C*) and cross-set size usually decreases upwards within such sandstone units. Planar cross-strata and horizontal lamination are subordinate.

Deposition was probably in a network of poorly confined braided-stream channels. Sandstone units vary in their thickness and facies content according to position on the alluvial plain. Between main fluvial tracts (profile 12 in Fig. 12), the proportion of flood-basin siltstones is slightly higher and there are more common solitary sets of planar cross-strata which represent transverse bars. These latter were probably not preserved within main fluvial tracts (profile 11) due to spasmodic lateral migration of deeper, longer-active channels or channel pools (Rust, 1972), apparently conducive to the development of dune-type bedforms (trough cross-stratified infills).

This braided-stream system was supplied, laterally, with sediment-laden discharge from a tributary alluvial fan (Fig. 12, inset). At Craig-yr-Ysfa, 5 km to the north, correlative sediments (Fig. 14) consist of interbedded conglomerates, pebbly sandstones, and sandstones, whose lenticular units are 50–200 cm thick and display mainly trough cross-stratification (sets up to 70 cm thick). Palaeocurrents are unidirectional, towards the W. Conglomerates attributed to debris-flows are subordinate, but present. The deposition was probably in a medial/lower segment of a large, stream-dominated alluvial fan. Stacked coarsening-upward sequences (Fig. 14) may represent progradation and/or later shifting of individual fan-lobes. It is likely that a number of small alluvial fans merged downslope to form a broad, coalescent fan-system (Fig. 12, inset). The direction of sediment dispersal swung from westerly to southerly as the fluvial system advanced farther into the basin and adjusted to basin-floor topography. A medial, more distant portion of the resultant basin-axis braidplain is thought to be represented by the fluvial sequences in profiles 11 and 12 (Fig. 12, see also inset map).

A thin, stratigraphically equivalent succession at Cwm Clorad Isaf (profile 13 in Fig. 15), 3 km to the SE (see map in Fig. 12), represents coastal termination of this braidplain and reflects interaction between fluvial and marine processes. Common swaley cross-stratification (Fig. 13*B*), large wave ripples (Fig. 13*D*) and horizons of gravel lag, all indicate a nearshore zone with high storm-wave energy. Narrow, sand-filled channels (2 m deep and up to 10 m wide) are cut into wave-reworked deposits and display either trough cross-stratification or large solitary sets of low-angle cross-strata (Fig. 13*A*). They represent prograding fluvial distributaries. The discharge from a network of braided streams on the lower delta plain was probably funnelled through discrete channels as the flow adjusted to shore-parallel barrier spits (Fig. 12, inset map).

The style of deposition on the delta front would thus have varied laterally, with simpler, coarsening-upward sequences developed off the main distributary channel-mouths, and with complex, barred-shoreline sequences formed between these active sites of fluvial clastic input. The Capel Curig successions then represents a wave-modified braidplain delta, where both stream effluent and wave action left distinct signatures in the delta-front sand prism.

A thick sequence of delta-retreat facies (phases 8–11 in Fig. 12) reflects gradual waning of fluvial sediment supply to the delta front and a concurrently increasing intensity of wave energy along the shoreline. The observed facies assemblages, their vertical arrangement and palaeocurrent pattern (Fig. 12) are all indicative of deposition in a transgressive, high-energy, barred shoreface environment (cf. Bourgeois, 1980; Leithold and Bourgeois, 1984).

Discussion

Sedimentation patterns and controlling factors

Palaeogeographical configurations varied between each of the five deltaic systems, with major changes in sediment dispersal following volcanic eruptions and emplacement of voluminous ashflow sheets (Orton, 1986). The contrasting patterns of delta-front sedimentation reviewed above can be considered as a consequence of the varying mode of interaction between sediment-laden alluvial systems and basinal marine processes (Fig. 16). Since tidal deposits are rare within the studied successions, and where present (Fig. 8, phase 5 in profile 4) suggest microtidal conditions, the spectrum of delta types primarily reflects a competition between fluvial and wave processes.

Nearshore wave energy must have been influenced by the size, shape, and depth of the marine basin. Water depths can be roughly estimated (Fig. 16) from the thicknesses of individual deltaic sequences (Klein, 1974), especially where the sediment supply exceeds basin subsidence and where rapid coastal progradation preserves entire delta-front prisms. In the present case, alluvial (delta-plain) facies appear to have capped all deltaic sequences. Several episodes of deltaic progradation, however, are thought to have been facilitated by concurrent volcanotectonic uplift (Orton, 1986). Regional tumescence preceding the eruption and emplacement of the voluminous Pitts Head Tuff (Figs. 1, 2) was particularly important, and the Cae'r Gors and Gwastadnant fan-delta sequences thus give substantially underestimated pre-tumescence water depths (see applied corrections in Fig. 16). The deltas appear to have prograded into a moderately deep marine basin (Fig. 16), where wave energy was moderately high to very high, probably depending on coastline orientation. Large wave ripples are present at all stratigraphical levels (e.g., profile 9 in Fig. 10; profile 13 in Fig. 15) and suggest that maximum wind fetch was from the present south/southeast. The slightly lower nearshore wave-energy confronting the Gwern Gof depositional system can thus be explained by the delta facing obliquely away fron the direction of main wave attack.

The inferred local variations in overall wave climate within the basin appear, however, to be rather insignificant when compared with the actual spectrum of delta types. The contrasting delta morphologies and associated patterns of sedimentation (summary in Fig. 16) are thus primarily a consequence of the differences in fluvial processes. The critical parameters which controlled these deltas and determined their differing styles of development are thought to have been the geometry and slope of the alluvial 'feeder' system, the rate of sediment supply (discharge magnitude and temporal variability), and the grain size of sediment delivered to the delta front. These physical parameters varied among and between the alluvial fans and braidplains, and dictated the varying responses of the associated deltas to basinal wave-climate.

When considering the processes of nearshore sedimentation on the delta front, it is useful to discriminate between dissipative and reflective shoreline morphologies (Wright *et al.*, 1979; Short and Hesp, 1981). Dissipative coastlines have low gradients, a complex topography with common longshore bars cut by rip currents, and waves breaking a considerable distance from the shoreline. In contrast, reflective coastlines are characterized by smooth nearshore profiles with most wave energy reflected directly from the beachface. Abundant fine sediment, a wide and shallow inshore profile, and high wave energy usually produce dissipative conditions; a coarser grain size, reduced sediment supply and lower wave power increase reflectiveness at the beachface.

	Depositional system	FLUVIAL REGIME: NATURE of FLUVIAL SYSTEM	FLUVIAL REGIME: SEDIMENT	DELTAIC RESPONSE: RATE OF SUPPLY	DELTAIC RESPONSE: N ↑, 1 km, sediment dispersal →	BASINAL REGIME: WAVE ENERGY	BASINAL REGIME: WATER DEPTH (deposits - based on)
FAN DELTAS	GWERN GOF	FIXED DISTRIBUTARY CHANNEL	fine to medium sand	HIGH	FLUVIALLY DOMINATED: DISTRIBUTARY CHANNELS, FLAT DELTA PLAIN	MODERATE TO HIGH	65m; profile 2 phase 1,2
	MOEL HEBOG	MOBILE BRAIDED STREAMS	pebbly coarse sand	MODERATE	FLUVIAL WAVE INTERACTION: REDEPOSITION OF FAN SALIENTS LATERALLY	HIGH	70m; profile 4, phase 1-3
	CAER GORS	UNCHANNELISED DEBRIS-FLOWS AND SHEETFLOODS	fine sand to gravel	LOW	WAVE DOMINATED: WAVE DOMINATED COASTLINE	HIGH	20m; profile 6 phase 2 & 3 65m; corrected for volcanotectonic uplift
	GWAS TADNANT	UNCHANNELISED DEBRIS-FLOWS AND SHEET FLOODS (inferred)	medium coarse sand to gravel	LOW	LITTLE PRESERVATION OF ALLUVIAL FACIES	EXTREMELY HIGH	25m; profile 8 phase 2 60m; corrected for volcanotectonic uplift
BRAIDPLAIN DELTAS	TRYFAN FACH	SMALL CHANNELS AND SHEETFLOODS MOBILE BRAIDED STREAMS	siliceous mud very coarse pebbly sand	EXTREMELY HIGH associated with period of volcanicity	FLUVIALLY DOMINATED: PROGRADATION AS SHEET, BARRED MUDDY SHELF	MODERATE	(~40m?)
	CAPEL CURIG	MOBILE BRAIDED STREAMS	medium to very coarse pebbly sand	MODERATE	FLUVIAL WAVE INTERACTION: BARRIER ISLANDS, FOCUS FLUVIAL INPUT, MOUTH BAR	EXTREMELY HIGH	75m; profile 11 phase 1-4

Figure 16. Salient features of the Caradocian fan deltas and braidplain deltas, North Wales. Sedimentology of the Tryfan Fach braidplain delta is not discussed and is to be published elsewhere (paper in preparation).

Fan deltas *v.* braidplain deltas

The contrasting depositional styles of fan deltas and braidplain deltas can be considered and interpreted against the background of the physical factors outlined above. This is best revealed by a comparison of two superficially similar, wave-modified deltaic systems: the Moel Hebog fan delta and the Capel Curig braidplain delta (Fig. 16). Both systems delivered sediment of similar size grade to the delta front by unstable braided streams, and both prograded into a basin with high nearshore energy. They differed, however, in their overall geometry, their nearshore evolution, and their ability to cope with marine energy.

The Moel Hebog fan delta was characterized by a lobate or arcuate geometry, partially inherited from its associated alluvial fan, and by a lower, more 'flashy' sediment supply, as the system relied solely on one drainage basin as a source. The common occurrence of debris-flow deposits suggests that the alluvial plain of this delta was of a steeper gradient. To some extent, it is the alluvial system that determines the inshore slope of the delta front, especially the part which is above modal wave-base. For example, Dunne and Hempton (1984) examined two contrasting styles of deltaic sedimentation in the pull-apart Lake Hazar basin and noted that an alluvial slope of at least 0.05(3°) was critical to produce fan deltas with steep, Gilbert-type foresets of 0.50(28°) slope. Lower-gradient braidplains in the axial zone of Lake Hazar basin resulted in fine-grained, low-angle (1°), 'mouth-bar-type' deltas (see also Wood and Ethridge, this volume).

In the Moel Hebog fan-delta system, dissipative conditions were associated with mouth bars; rip-current deposits resulted from wave-induced coastal set-up (cf. Wright, 1977; Wright *et al.*, 1980). However, a combination of smaller overall sediment input and steeper inshore slope on the fan delta resulted in reflective beachface conditions characterized by a simple, two-dimensional, coarsening-upward shoreface profile along much of the delta front. With waves dispensing energy directly on the beachface due to steeper gradients and with a general deficit of inshore sediment storage, a reflective shoreline is likely to experience greater erosion on the beachface (Wright *et al.*, 1979).

The arcuate geometry of fan-delta fronts further assists wave processes in the redistribution of delta-front sediments. Wave refraction has two effects. By reducing wave height, it enhances the development of reflective beach morphologies, and by reducing the obliquity of wave incidence, it concentrates wave power on the fan salients. The curvature of the delta front will also induce longshore drift regardless of the angle of incidence of dominant wave-attack to overall shoreline orientation. A good modern example of such drift patterns is provided by the Jaba delta in Papua New Guinea, where normal wave incidence results in bidirectional sediment movement on either side of an axial river mouth (Wright *et al.*, 1980). In the Moel Hebog fan delta, the sediment from such longshore drift is inferred to have been distributed and accumulated laterally between fan salients as spit barriers. Due to this pattern of sediment dispersal, delta progradation was comparatively ineffective; depositional sequences farther to the NE (Fig. 8, inset map) are silt-dominated.

In contrast, the Capel Curig braidplain delta (Fig. 12) was probably fed laterally by alluvial fans, one of them being represented at Craig-yr-Ysfa. A combination of higher sediment supply and lower inshore slope would result in the fully dissipative, barred coastline which has been inferred from facies sequences. Although dissipative shorelines have high rates of seaward sediment transport (Wright *et al.*, 1979), their large inshore sediment storage creates the greatest potential for net subaerial accretion. With any reduction (for instance seasonal) in incident wave height, much sediment will be returned to the shoreface (e.g. Clifton *et al.*, 1971) by migration and welding of nearshore bars and ridges (e.g. Hine and Boothroyd, 1978). Although such bars will influence the location of fluvial input, as in the Capel Curig example, they will not inhibit net progradation of the delta as a sheet-like prism.

Good modern analogues for the contrasting Moel Hebog and Capel Curig deltas are perhaps the Yallahs fan delta of Jamaica and the Skeidararsandur coastal braidplain of Iceland (Fig. 17). Both these depositional systems have seasonal discharge in a network of gravel-bed braided streams and prograde into marine basins with high wave energy and insignificant tidal influence on the pattern of sedimentation. In the Yallahs example, owing to the coarse size of sediment, limited sediment supply and steep inshore slope, waves expend much of their energy directly against the shoreface. Thus, most of the delta front is of erosive nature, with steep, coarse-grained foreshore zone.

In contrast, a much higher wave energy, a much greater sediment supply and a lower inshore slope have caused the beach and nearshore zones of the Skeidararsandur delta to develop a complex rhythmic topography that includes coastal barrier islands and large offshore bars. Dissipative nearshore conditions, combined with an oblique incidence of maximum wave energy, result in substantial seaward and longshore sediment transport. In spite of this, a 75-km stretch of the shoreline was able to advance 1.1 km in 300 years (Hine and Boothroyd, 1978). Progradation of this depositional system could produce patterns of delta-front sedimentation similar to the Capel Curig braidplain-delta example (Figs. 12, 15), with mouth bars reflecting localized sediment input through distributary channels and with shore-parallel bars and shoals characterizing the rest of the delta front.

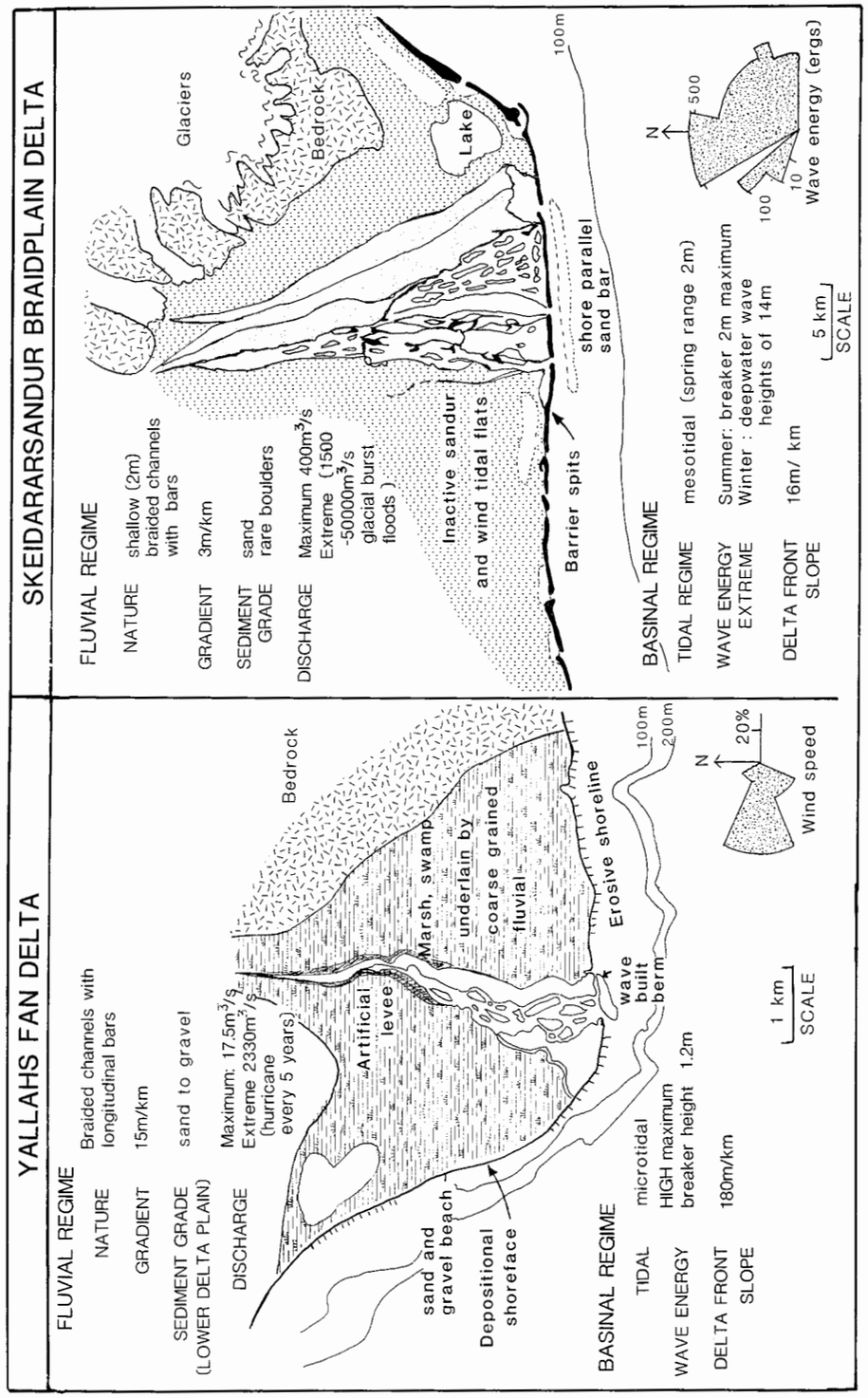

Figure 17. Comparison of the Yallahs fan delta, Jamaica, and the Skeidararsandur (braidplain) delta, Iceland. Data from Wescott and Ethridge (1980); Hine and Boothroyd (1978); Ward *et al.* (1976). Note scale difference.

Fan deltas: the importance of alluvial regime

Fan deltas are the result of progradation of alluvial fans into standing bodies of water. The size and morphology of fan deltas are largely dictated by the nature of the contributing alluvial fan. Modern alluvial fans can be divided into two end-member categories: fluvially dominated and mass-flow dominated (Bull, 1962, 1972). Fluvially dominated fans are of larger size and lower gradient, and generally display good correlation between their gradients and the size of transported sediment (e.g. Boothroyd, 1972; Schramm, 1981). This relationship is useful in interpreting ancient depositional sequences. Fans dominated by sediment gravity flows are smaller and have steeper slopes (Bull, 1962, 1972).

Fan morphology and drainage basin area are directly related (Denny, 1967; Hooke, 1967, 1968; Ryder, 1971; Harvey, 1984). With large catchment areas, more regular discharge results in large, fluvially dominated fans. With small, rugged drainage basins, where streams are ephemeral, fluvial processes are less effective in moving sediment, and hence mass-flows become more important. In addition, the steeper gradients in smaller basins render the sediment stored on slopes susceptible to instability and gravity-induced mass movements during heavy rainfalls (Wasson, 1977). Sediment movement on associated fans is infrequent and, depending on climatic conditions, may occur only during flood events every few centuries (Wasson, 1977; Kostaschuk *et al.*, 1986).

In the Devonian Hornelen Basin of Norway, small, steep, mass-flow dominated fans developed along the northern margin can be distinguished from large, fluvially dominated fans of the southern margin (Gloppen and Steel, 1981). The differences in fan thickness, morphology and facies are related to drainage-basin areas, with small alluvial catchments along the tectonically more active northern margin (Steel, 1976). Since the northern fans prograded into a shallow, low-energy floodbasin lake (Steel and Aasheim, 1978; Pollard *et al.*, 1982), the dominant influence of this passive body of standing water was subaqueous dilution of debris flows and a variety of 'gravity-winnowing' processes on water-saturated, steep subaqueous slopes (Larsen and Steel, 1978; Nemec *et al.*, 1984).

In the North Wales basin, in the absence of any known variations in climate or bedrock lithology, the recognized diversity of alluvial systems and resultant delta types is also attributed to the tectonic control on catchment areas. Although the deltas prograded into a basin with high wave energy, their deposits reveal that the nature of the fluvial input was crucial in dictating the effectiveness of the delta in coping with marine energy, and in controlling the extent of sediment reworking by waves and the type of resultant coastline.

The Gwern Gof fan delta was associated with channellized, low-sinuosity streams carrying an abundant bedload of fine/medium sand. Sediment composition and grain size suggest deposition during a tectonically 'quiescent', non-volcanic period. The high rates of sediment supply, relative to the effective work of sea waves in the nearshore zone, enabled multiple progradation of large, fluvially dominated delta (Fig. 5). The comparatively fine sediment grain size and inherent low-gradient slope of the delta promoted progradation in a dissipating wave-energy regime. Interdistributary bays and coastal lagoons were important components of the flat lower delta-plain.

The wave-dominated Gwastadnant and Cae'r Gors fan deltas represent the other end of the spectrum in this study. These were associated with small alluvial fans dominated by sediment gravity flows. Development of any delta-plain lagoons was precluded by the steep (*c.* 5°) fan gradients. In a study of modern, coarse-grained, stream-dominated fan deltas on a microtidal, low wave-energy coast of Honduras, Schramm (1981) similarly noted the control of fan slope on lower delta-plain environments, with the delta plain consisting of longitudinal bars on small fan deltas. In the Welsh examples, the progradational delta-plain sequences comprise 'proximal' debris-flow conglomerates overlying sandy sheetflood deposits, suggesting downfan dilution of sediment flows and decrease in their competence. The extreme variation in grain size delivered to the delta front resulted in the spectrum of coastline morphologies discussed above. Low sediment supply, coarser grain size and steeper inshore gradient, however, favoured development of a fully reflective shoreline in the Cae'r Gors and Gwastadnant examples. As a result of infrequent sediment input, which must have characterized these fans, wave processes were able to redistribute most of the sediment delivered to the delta front, with dispersal both offshore and alongshore. The finger-grained, relatively passive, distal portions of mass flow-dominated alluvial fans would be particularly susceptible to marine erosion and redeposition. The preservation of the thin 'proximal' alluvial sequences in the Cae'r Gors and Gwastadnant fan-delta examples may be, in fact, only a consequence of the early burial by emplacement of the overlying Pitts Head Tuff (Figs. 1, 2).

Towards a classification

When interpreting ancient coarse-grained deltas, one should first attempt to discriminate between fan deltas, braidplain deltas and braid deltas (as defined here). The examples from North Wales reveal the importance of alluvial system type in determining: (i) the effectiveness of discharge in coping with marine energy; (ii) the pattern of delta-front sedimentation; and (iii) the development of either reflective or dissipative shorelines. Several authors (e.g. Rust and Koster, 1984, p. 164; McPherson *et al.*, 1987) have also discussed the contrasting tectonic implications of these various types of alluvial systems.

In fan deltas, the size grade and abundance of sediment will be a critical factor in governing coastal response to wave energy and resultant patterns of delta-front sedimentation. Sediment grain-size will also strongly influence the gradient of the delta front. It is not surprising that most examples of small fan deltas, especially those on which sediment gravity flows are the dominant alluvial process, have been described from lacustrine settings or from very small basins where subaqueous slopes are not significantly re-

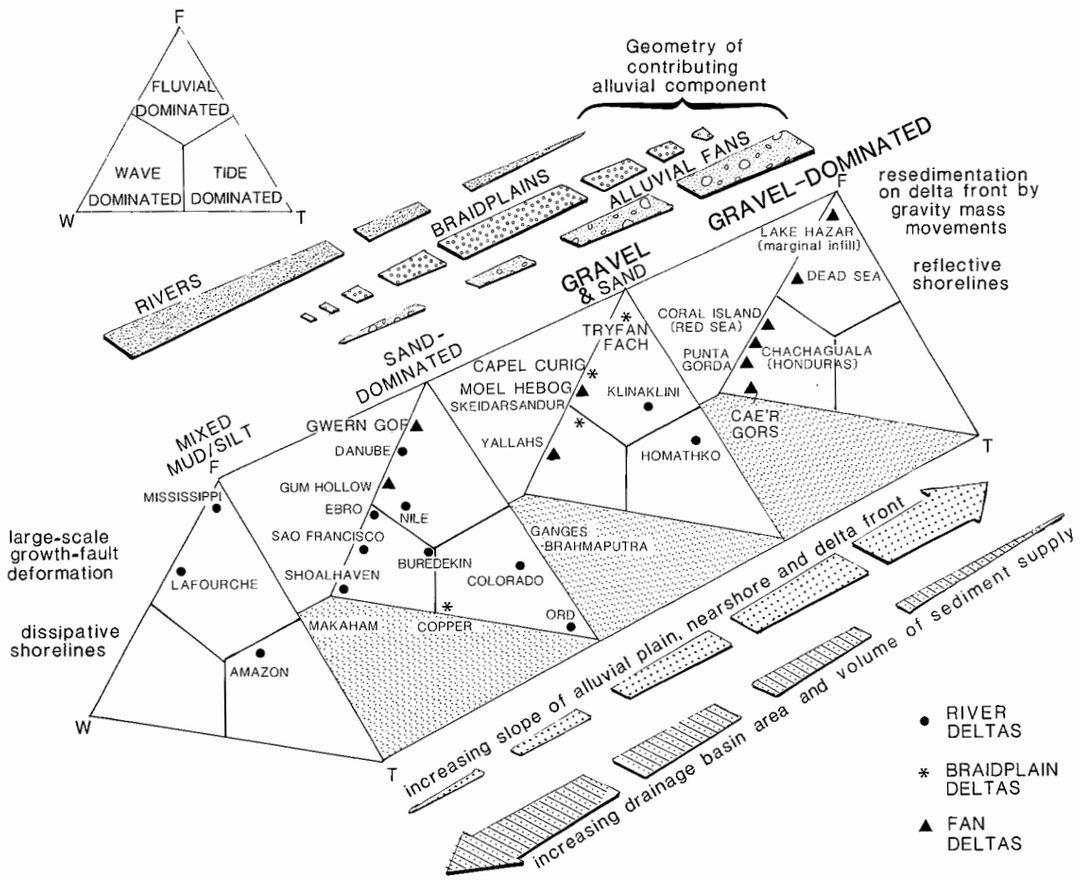

Figure 18. Suggested classification of deltas, based on the dominant processes (fluvial, wave, or tidal) of sediment dispersal at the delta front and on the prevailing grain size of sediment delivered to the front. Grain size is used to reflect relative sizes of drainage basins, although this criterion may not apply to alluvial systems affected by active volcanicity or dominated by mass-flows. Tidal influence will be minor in coarse-grained deltas because of the relatively steep slope of their alluvial plains. Ancient delta examples are from this study; modern examples are from the literature (Hine and Boothroyd, 1978; Sneh, 1979; Wescott and Ethridge, 1980; Schramm, 1981; Hayes and Michel, 1982; Hayward, 1985; Wright, 1985).

worked by wave processes (e.g. Larsen and Steel, 1978; Postma, 1983; Nemec *et al.*, 1984). Even where protected by patch reefs, the low rate of sediment supply and coarse sediment size allow development of erosive reflective shorelines (Hayward, 1985; Dabrio and Polo, this volume). The inherent steep slopes in such fan deltas will also promote subaqueous redistribution of delta-front deposits by slumping and gravity mass-flow (e.g. Postma, 1983, 1984). Deltaic cyclicity will often result from progradation of the entire alluvial fan, rather than from deposition related to seaward decrease in discharge momentum and grain size of sediment transported by channel effluent (as in 'ordinary' deltas). At the other end of the spectrum, large alluvial fans, owing to their finer-grained and more abundant supply of sediment, will be more efficient in coping with basin energy. Large-scale growth faulting will be the dominant deformation feature on the delta front.

The morphology, sediment distribution and facies patterns of both fan and braidplain deltas is controlled by competing fluvial, tidal, and wave processes. It is reasonable, therefore, that deposits should be classified according to the dominant process (Kleinspehn *et al.*, 1984), as in the case of 'ordinary' river deltas (Galloway, 1975). However, to better understand coarse-grained deltas, and perhaps all delta types (Reading, 1986, p. 521), the ternary classification diagram of common deltas should be extended in a fourth dimension to account for the grain size of sediment delivered to the delta front along the axis of the alluvio-deltaic prism (Fig. 18). A similar approach has recently been used to comprehend shallow-marine siliciclastic facies (Johnson and Baldwin, 1986, p. 258). This makes it possible to compare deltas with similar plain/front gradients, similar responses to nearshore wave energy, and similar catchment areas and rates of clastic supply. The various types of alluvial systems will overlap along the 'sediment grade' axis, or the fourth dimension, of the extended classification diagram (Fig. 18).

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References

Boothroyd, J.C. (1972) Coarse-grained sedimentation on a braided outwash fan, Northeast Gulf of Alaska. *Tech. Rept.* **6**, Coastal Res. Div., University of South Carolina, 127 pp.

Boothroyd, J.C. and Ashley, G.M. (1975) Processes, bar morphology and sedimentary structures on braided outwash fans, NE Gulf of Alaska. In Glaciofluvial and Glaciolacustrine Sedimentation, eds. A.V. Jopling and B.C. McDonald, *Spec. Publ. Soc. econ. Paleontol. Mineral.* **23**, 193–222.

Boothroyd, J.C. and Nummedal, D. (1978) Proglacial braided outwash: a model for humid alluvial fan deposits. In Fluvial Sedimentology, ed. A.D. Miall, *Mem. Can. Soc. Petrol. Geol.* **5**, 641–668.

Bourgeois, J. (1980) A transgressive shelf sequence exhibiting hummocky cross stratification: The Cape Sebastian sandstone (Upper Cretaceous), SW Oregon. *J. sediment. Petrol.* **50**, 681–702.

Bull, W.R. (1962) Relation of alluvial fan size and slope to drainage basin size and lithology in western Fresno County, California. *Prof. Pap. U.S. geol. Surv.* **450B**, 51–53.

Bull, W.R. (1972) Recognition of alluvial fan deposits in the stratigraphic record. In Recognition of Ancient Sedimentary Environments, eds. J.K. Rigby and W.K. Hamblin, *Spec. Publ. Soc. econ. Paleontol. Mineral.* **16**, 63–83.

Campbell, C.V. (1976) Reservoir geometry of a fluvial sheet sandstone. *Bull. Am. Assoc. Petrol. Geol.* **70**, 1009–1020.

Clifton, H.E. (1981) Progradational sequences in Miocene shoreline deposits, southeastern Caliente Range, California. *J. sediment. Petrol.* **51**, 165–184.

Clifton, H.E., Hunter, R.E. and Phillips, R.L. (1971) Depositional structures and processes in the high energy nearshore. *J. sediment. Petrol.* **41**, 651–670.

Denny, C. (1967) Fans and pediments. *Am. J. Sci.* **265**, 81–105.

Dixon, J. (1979) The Lower Cretaceous Atkinson Point Formation (new name) on the Tuktoyakaktuk peninsula, N.W.T.: A coastal fan delta to marine sequence. *Bull. Can. Pet. Geol.* **27**, 163–182.

Dunne, L.A. and Hempton, M.R. (1984) Deltaic sedimentation in the Lake Hazar pull-apart basin, southeastern Turkey. *Sedimentology* **31**, 401–412.

Elliott, T. (1974) Interdistributary bay sequences and their genesis. *Sedimentology* **21**, 611–622.

Ethridge, F. G. and Wescott, W.A. (1984) Tectonic setting, recognition, and hydrocarbon reservoir potential of fan delta deposits. In Sedimentology of Gravels and Conglomerates, eds. E.H. Koster and R.J. Steel, *Mem. Can. Soc. Petrol. Geol.* **10**, 217–235.

Farquharson, G.W. (1982) Lacustrine deltas in a Mesozoic alluvial sequence from Camp Hill, Antarctica. *Sedimentology* **29**, 717–725.

Flores, R.M. (1975) Short-headed stream delta: model for Pennsylvanian Haymond Formation. *Bull. Am. Assoc. Petrol. Geol.* **59**, 2288–2301.

Flores, R.M. and Ferm, J.C. (1970) A recent model for Pennsylvanian deposition in the Marathon Basin, West Texas. *J. sediment. Petrol.* **40**, 621–628.

Galloway, W.E. (1975) Process framework for describing the

morphological and stratigraphic evolution of deltaic depositional systems. In *Deltas, Models for Exploration*, ed. M.L. Broussard, Houston Geological Society, 87–98.

Galloway, W. E. (1976) Sediments and stratigraphic framework of the Copper River fan-delta, Alaska. *J. sediment. Petrol.* **46**, 726–737.

Galloway, W.E. (1981) Depositional architecture of Cenozoic Gulf Coastal plain fluvial systems. In Recent and Ancient Non-Marine Depositional Environments: Models for Exploration, eds. F.G. Ethridge and R.M. Flores, *Spec. Publ. Soc. econ. Paleontol. Mineral.* **31**, 127–155.

Galloway, W.E. and Hobday, D.K. (1983) *Terrigenous Clastic Depositional Systems: Applications to Petroleum, Coal and Uranium Exploration*. Springer, New York, 423 pp.

Gloppen, T.G. and Steel, R.J. (1981) The deposits, internal structure and geometry in six alluvial fan-delta bodies (Devonian, Norway) – a study in the significance of bedding sequences in conglomerates. In Recent and Ancient Non-Marine Depositional Environments: Models for Exploration, eds. F.G. Ethridge and R.M. Flores, *Spec. Publ. Soc. econ. Paleontol. Mineral.* **31**, 49–69.

Harbaugh, D.W. and Dickinson, W.R. (1981) Depositional facies of Mississippian clastics, Antler foreland basin, Central Diamond Mountains, Nevada, *J. sediment. Petrol.* **51**, 1223–1234.

Harvey, A.M. (1984) Debris flows and fluvial deposits in Spanish Quaternary alluvial fans: implications for fan morphology. In Sedimentology of Gravels and Conglomerates, eds. E.H. Koster and R.J. Steel, *Mem. Can. Soc. Petrol. Geol.* **10**, 123–132.

Hayes, M.O. and Michel, J. (1982) Shoreline sedimentation within a forearc embayment, Lower Cook Inlet, Alaska. *J. sediment. Petrol.* **52**, 251–263.

Hayward, A.B. (1985) Coastal alluvial fans and associated marine facies in the Miocene of S.W. Turkey. In Modern and Ancient Fluvial Systems, eds. J.D. Collinson and J. Lewin, *Spec. Publ. Int. Assoc. Sedimentol.* 323–336.

Heward, A.P. (1981) A review of wave dominated clastic shoreline deposits. *Earth Sci. Rev.* **17**, 223–276.

Hine, A.C. and Boothroyd, J.C. (1978) Morphology, processes and sedimentary history of a glacial outwash plain shoreline, southern Iceland. *J. sediment. Petrol.* **48**, 901–920.

Hobday, D.K. (1974) Interactions between fluvial and marine processes in the lower part of the Late Precambrian Vadsø Group, Finnmark. *Nor. geol. Unders.* **303**, 39–56.

Hobday, D.K. and von Brunn, V. (1979) Fluvial sedimentation and palaeography of an early Paleozoic failed rift, southeastern margin of Africa. *Palaeogeogr. Palaeoclimatol. Palaeoecol.* **28**, 169–184.

Holmes, A. (1965) *Principles of Physical Geology*. 2nd edn., Roland Press, New York, 1288 pp.

Hooke, R. LeB. (1967) Processes in arid region alluvial fans. *J. Geol.* **75**, 438–460.

Hooke, R. LeB. (1968) Steady state relationships on arid region alluvial fans. *J. Geol. (Chicago)* **75**, 438–460.

Howard, J.D. and Reineck, H.E. (1981) Depositional facies of high energy beach to offshore sequences: comparison with low energy sequence. *Bull. Am. Assoc. Petrol. Geol.* **65**, 807–830.

Howells, M.F., Leveridge, B.F., Addison, R. and Reedman, A.J. (1983) The lithostratigraphic subdivision of the Ordovician underlying the Snowdon and Crafnant volcanic groups, North Wales. *Rept. Inst. geol. Sci.* **83/1**, 11–15.

Hunter, R.E., Clifton, H.E. and Phillips, R.L. (1979) Depositional processes, sedimentary structures, and predicted vertical sequence in barred nearshore systems, Southern Oregon coast. *J. sediment. Petrol.* **49**, 711–726.

Johnson, H.D. and Baldwin, C.T. (1986) Shallow siliciclastic seas. In *Sedimentary Environments and Facies*, ed. H.G. Reading, Blackwell Scientific, Oxford, 229–282.

Kingsley, C.S. (1984) Dagbreek fan delta: An alluvial placer to prodelta sequence in the Proterozoic Welkom Goldfield, Witwatersrand, South Africa. In Sedimentology of Gravels and Conglomerates, eds. E.H. Koster and R.J. Steel, *Mem. Can. Soc. Petrol. Geol.* **10**, 321–330.

Klein, G. DeV. (1974) Estimates of water depths from analysis of barrier islands and deltaic sedimentary sequences. *Geology* **2**, 409–412.

Kleinspehn, K.L., Steel, R.J., Johannessen, E. and Netland, A. (1984) Conglomeratic fan-delta sequences, Late Carboniferous – Early Permian, Western Spitsbergen. In Sedimentology of Gravels and Conglomerates, eds. E.H. Koster and R.J. Steel, *Mem. Can. Soc. Petrol. Geol.* **10**, 279–294.

Kokelaar, B.P., Howells, M.F., Bevins, R.E., Roach, R.A. and Dunkley, P.N. (1984) The Ordovician marginal basin of Wales. In Marginal Basin Geology: Volcanic and Associated Sedimentary and Tectonic processes in Modern and Ancient Marginal Basins, eds. B.P. Kokelaar and M.F. Howells, *Spec. Publ. geol. Soc. London* **16**, 245–279.

Kostachuk, R.A., MacDonald, G.M. and Putnam, P.E. (1986) Depositional process and alluvial fans – drainage basin morphometric relationships near Banff, Alberta, Canada. *Earth Surf. Proc.* **11**, 471–484.

Larsen, V. and Steel, R.J. (1978) The sedimentary history of a debris-flow dominated, Devonian alluvial fan – a study of textural inversion. *Sedimentology* **25**, 37—59.

Leithold, E.H. and Bourgeois, J. (1984) Characteristics of coarse-grained sequences deposited in nearshore wave dominated environments – examples from the Miocene of SW Oregon. *Sedimentology* **25**, 37–59.

McGowen, J.H. (1971) Gum Hollow fan delta, Nueces Bay, Texas. *Rep. Invest. Bur. econ. Geol. Univ. Texas* **69**, 91 pp.

McGowen, J.H. and Scott, A.J. (1974) Fan delta deposition: processes, facies and stratigraphic analogues. *Abstracts, Am. Assoc. Petrol. Geol. Soc. econ. Paleontol. Mineral.* pp. 60–61.

McPherson, J.G., Shanmugam, G. and Miola, R.J. (1987) Fan-deltas and braid deltas: varieties of coarse-grained deltas. *Bull. geol. Soc. Am.* **99**, 331–340.

Massari, F. (1984) Resedimented conglomerates of a Miocene fan delta complex, Southern Alps, Italy. In Sedimentology of Gravels and Conglomerates, eds. E.H. Koster and R.J. Steel, *Mem. Can. Soc. Petrol. Geol.* **10**, 259–278.

Miall, A.D. (1977) A review of the braided river depositional environment. *Earth Sci. Rev.* **13**, 1–16.

Moore, G.T. and Asquith, D.O. (1971) Delta: term and concept. *Bull. geol. Soc. Am.* **82**, 2563–2568.

Nemec, W. and Steel, R.J. (1984) Alluvial and coastal conglomerates: their significant features and some comments on gravelly mass-flow deposits. In Sedimentology of Gravels and Conglomerates, eds. E.H. Koster and R.J. Steel, *Mem. Can. Soc. Petrol. Geol.* **10**, 1–33.

Nemec, W., Steel, R.J., Porębski, S.J. and Spinnangr, Å. (1984)

Domba Conglomerate, Devonian, Norway: process and lateral variability in a mass flow-dominated, lacustrine fan-delta. In Sedimentology of Gravels and Conglomerates, eds. E.H. Koster and R.J. Steel, *Mem. Can. Soc. Petrol. Geol.* **10**, 295–320.

Orton, G.J. (1986) Tectonic control of mid-Ordovician volcaniclastic sedimentation, North Wales (Abstract). *Bull. Am. Assoc. Petrol. Geol.* **70**, 628.

Pollard, J.E., Steel, R.J. and Undersrud, E. (1982) Facies sequences and trace fossils in lacustrine/fan delta deposits, Hornelen Basin (Middle Devonian), western Norway. *Sediment. Geol.* **32**, 63–87.

Postma, G. (1983) Water escape structures in the context of a depositional model of a mass-flow dominated conglomeratic fan-delta (Abrioja Formation, Pliocene, Almeria Basin, SE Spain). *Sedimentology* **30**, 91–103.

Postma, G. (1984) Mass-flow conglomerates in a submarine canyon: Abrioja fan-delta, Pliocene, southeast Spain. In Sedimentology of Gravels and Conglomerates, eds. E.H. Koster and R.J. Steel, *Mem. Can. Soc. Petrol. Geol.* **10**, 237–258.

Prior, D.B., Bornhold, B.D. and Johns, M.W. (1986) Active sand transport along a fjord bottom channel, Bute Inlet, British Columbia. *Geology* **14**, 581–584.

Pudsey, C.J. (1984) Fluvial to marine transition in the Ordovician of Ireland – a humid-region fan-delta? *Geol. J.* **19**, 143–172.

Reading, H.G. (1986) Problems and perspectives. In *Sedimentary Environments and Facies*, ed. H.G. Reading, Blackwell Scientific, Oxford, 520–524.

Reedman, A.J., Colman, T.B., Campbell, S.D.G. and Howells, M.F. (1985) Volcanogenic mineralization related to the Snowdon Volcanic Group. *J. geol. Soc. London* **142**, 875–888.

Reid, I. and Frostick, L.E. (1985) Beach orientation, bar morphology and the concentration of metalliferous placer deposits: a case study, Lake Turkana, N. Kenya. *J. geol. Soc. London* **142**, 837–848.

Rust, B.R. (1972) Structures and process in a braided river. *Sedimentology* **18**, 221–245.

Rust, B.R. (1979) Coarse alluvial deposits. In *Facies Models*, ed. R.G. Walker, *Geosci. Can. Reprint Ser.* **1**, 53–70.

Rust, B.R. and Koster, E.H. (1984) Coarse alluvial deposits. In *Facies Models*, 2nd edn., ed. R.G. Walker. *Geosci. Can. Reprint Ser.* **1**, 53–70.

Ryder, J.M. (1971) Some aspects of the morphometry of paraglacial alluvial fans in south-central British Columbia. *Can. J. Earth Sci.* **8**, 1252–1264.

Schramm, W.E. (1981) Humid Tropical Alluvial Fans, Northwest Honduras. Unpubl. M.Sc. thesis, Louisiana State University.

Short, A.D. and Hesp, P.A. (1981) Wave beach and dune interactions in southeastern Australia. *Marine Geol.* **48**, 259–284.

Smith, N.D. (1972) Some sedimentary aspects of planar cross-stratification in a sandy braided river. *J. sediment. Petrol.* **42**, 624–634.

Sneh, A. (1979) Late Pleistocene fan deltas along the Dead Sea Rift. *J. sediment. Petrol.* **49**, 541–552.

Steel, R.J. (1976) Devonian basins of western Norway: sedimentary response to tectonism and varying tectonic context. *Tectonophysics* **36**. 207–224.

Steel, R.J. and Aasheim, S.M. (1978) Alluvial sand deposition in a rapidly subsiding basin (Devonian, Norway). In Fluvial Sedimentology, ed. A.D. Miall, *Mem. Can. Soc. Petrol. Geol.* **5**, 386–412.

Surlyk, F. (1984) Fan-delta to submarine fan conglomerates of the Volgian-Valanginian Wollaston foreland group, East Greenland. In Sedimentology of Gravels and Conglomerates, eds. E.H. Koster and R.J. Steel, *Mem. Can. Soc. Petrol. Geol.* **10**, 359–382.

Sykes, R.M. and Brand, R.P. (1976) Fan delta sedimentation: an example from the late Jurassic-early Cretaceous of Milneland, Central East Greenland. *Geol. Mijnbouw* **55**, 195–203.

Syvitski, J.P.M. and Farrow, G.F. (1983) Structures and processes in bayhead deltas: Knight and Bute Inlet, British Columbia. *Sediment. Geol.* **36**, 217–244.

Van der Meulen, S. (1983) Internal structure and environmental reconstruction of Eocene transitional fan delta deposits, Monilobat-Castigaleu formations, Southern Pyrenees, Spain. *Sediment. Geol.* **37**, 85–112.

Vos, R.G. (1981) Sedimentology of an Ordovician fan delta complex, western Libya. *Sediment. Geol.* **29**, 153–170.

Vos, R.G. and Eriksson, K.A. (1977) An embayment model for tidal and wave swash deposits occurring within a fluvial dominated middle Proterozoic sequence in South Africa. *Sediment. Geol.* **18**, 161–173.

Ward, L.G., Stephen, M.F. and Nummedal, D. (1976) Hydraulics and morphology of glacial outwash distributaries, Skeidararsandur, Iceland. *J. sediment. Petrol.* **46**, 770–777.

Wasson, P.J. (1977) Catchment processes and the evolution of alluvial fans in the lower Derwent valley, Tasmania. *Z. Geomorphol.* **21**, 147–168.

Wescott, W.A. and Ethridge, F.G. (1980) Fan delta sedimentology and tectonic setting–Yallahs fan delta, southeast Jamaica. *Bull. Am. Assoc. Petrol. Geol.* **64**, 374–399.

Wescott, W.A. and Ethridge, F.G. (1982) Bathymetry and sediment dispersal dynamics along the Yallahs fan delta front, Jamaica. *Marine Geol.* **46**, 245–260.

Wright, L.D. (1977) Sediment transport and deposition at river mouths: a synthesis. *Bull. geol. Soc. Am.* **88**, 857–868.

Wright, L.D. (1985) River deltas. In *Coastal Sedimentary Environments*, ed. R.A. Davis, Springer, New York, 1–76.

Wright, L.D., Chappel, I., Thom, B.G., Bradshaw, M.P. and Cowell, M.P. (1979) Morphodynamics of reflective and dissipative beach and inshore systems, southeastern Australia. *Marine Geol.* **32**, 105–140.

Wright, L.D., Thom, B.G. and Higgins, R.J. (1980) Wave influences on river mouth depositional processes: examples from Australia and Papua, New Guinea. *Estuar. Coast. Marine Sci.* **11**, 263–277.

Fan Deltas: Sedimentology and Tectonic Settings
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Pliocene–Holocene fan deltas and braid deltas in the Crati Basin, southern Italy: a consequence of varying tectonic conditions

ALBINA COLELLA

The Pliocene–Holocene Crati Basin was formed due to E–W crustal extension related to the opening of the Tyrrhenian Sea. The basin comprises three sectors, each of which has had a different tectono-sedimentary evolution: (i) the Crati Trough is of extensional origin; (ii) the Cassano Domain is a shear zone along the main strike-slip Pollino fault system; and (iii) the Corigliano Trough is a downthrown basement block, now the subaqueous part of the basin. The Crati Basin has been filled by deltas and fan deltas with differing architecture, size, geometry, and petrology. Braid deltas as well as shelf-type and Gilbert-type fan deltas have developed in response to varying structural conditions in the basin.

The modern Crati River is a braided system flowing along the axis of the basin. On entering the Ionian Sea, the river forms a braid delta which is rapidly prograding on to a deep-water slope and feeding a mud-rich turbidite fan. This composite, fluviodeltaic-turbiditic system derives sediment through cannibalism of the elevated Pliocene–Holocene deposits. Shelf-type fan deltas developed along the margins of the Crati Basin mainly during the late Pliocene-early Pleistocene, due to first rifting stage, when an overall subsidence of the basin floor occurred.

Nine Gilbert-type fan-delta complexes occur in the Crati Basin. In the Cassano Domain, they developed during the early Pleistocene as vertically stacked wedges, in response to repetitive, rapid vertical movement along the sinistral Pollino Line. In the Crati Trough, Gilbert-type fan deltas replaced shelf-type fan deltas at the end of early Pleistocene, when a rapid differential uplift affected both the basin margins and its interior; this caused segmentation of the basin into narrow compartments separated by intrabasinal structural highs. At this stage, vertical displacements of up to some metres occurred on the individual faults. This caused the development of steep, fault-bounded coastlines with Gilbert-type fan deltas. The progradation of these latter systems was favoured by (i) the rapid uplift of the drainage areas: (ii) a small amount of basin subsidence; (iii) narrow coastlines; (iv) an abundant coarse sediment supply to the coastlines that fringed broader submarine platforms; and (v) the configuration of the basin, which protected it from strong longshore and tidal currents.

The occurrences of Gilbert-type fan deltas in the Crati Basin, as well as in other ancient and modern basins of the Mediterranean area, indicate that the development of such systems is not restricted to intracratonic basins, as suggested by Ethridge and Wescott (1984), but is more generally related to strike-slip and small extensional basins, regardless of their broader tectonic setting.

Introduction

The Crati Basin (Pliocene–Holocene), southern Italy, contains three different types of deltaic system: (i) a large braid delta (*sensu* McPherson *et al.*, 1987) which feeds a deep-water turbidite fan; (ii) shelf-type fan deltas (*sensu* Wescott and Ethridge, 1980), characterized by a very low-gradient offshore profile; and (iii) Gilbert-type fan deltas. Tectonics was the dominant control on sedimentation during the filling of this extensional basin, and the aim of this chapter is to discuss how changes in fault slip-rate and structural configuration have resulted in varied architecture of deltaic bodies, particularly of Gilbert-type fan deltas.

Gilbert-type deltas are highly constructive systems which were formerly assumed to develop preferentially in lacustrine basins (Gilbert, 1885, 1890; Gustavson *et al.*, 1975; Stanley and Surdam, 1978; McLaughlin and Nilsen, 1982; Dunne and Hempton, 1984). This assumption arose largely from the realization that a homopycnal inflow (Bates, 1953) would develop in such setting, and that the rapid mixing of fluvial and basinal waters would favour sediment deposition close to the river mouth.

In recent years, however, there has been an increasing recognition of the occurrence of recent and ancient Gilbert-type sequences in marine settings (Casey and Scott, 1979; Walker and Harms, 1980; Prior *et al.*,

ENCLOSURE 4:

Orton, G.J. (1988) A spectrum of fan and braidplain deltaic sequences, North Wales: a consequence of varying fluvial clastic input.

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